Assessment of soil water movement and the relative importance of shallow subsurface flow in a near-level Prairie watershed

by

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Abstract

Near-level Prairie landscapes have received limited attention in hydrological research. For this thesis, hydrometric measurements and four tracing experiments were completed at three “riparian-to-stream” sites in the Catfish Creek Watershed (southeastern Manitoba) to enhance Prairie hydrology understanding. First, hydrologic state variables were examined to infer vertical and lateral water movement. Second, tracer data were analyzed to evaluate the relative importance of surface versus subsurface water movement. Results show that hydrologic state variables can be useful for inferring riparian-to-stream water movement. Tracer data also revealed that subsurface water movement can contribute significantly to streamflow during snowmelt- and rainfall-triggered events in the study watershed. This thesis demonstrated that subsurface flow is a significant runoff generation mechanism in Prairie landscapes, thus challenging surface water-focused conceptualizations and management strategies that are traditionally used. The findings summarized in this thesis will be critical to improve the performance of hydrological models when applied to the Prairies.

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CHAPTER 1.

INTRODUCTION

1.1 General introduction

The discipline of hydrology pursues understanding of the occurrence, circulation and distribution of the earth's water and its atmosphere (Davie, 2008; Dingman, 2015). The hydrological cycle is defined by three primary components: runoff, evaporation and precipitation (Davie, 2008). "Runoff generation" is a phrase that encompasses the mechanisms via which precipitation is transformed into runoff and then routed from hillslopes to stream channels via various flow paths (Kirkby, 1988; Dingman, 2015). Several mechanisms of hillslope runoff generation have been well documented in the literature (Beven, 2006), although some of them are easier to measure in the field than others. Runoff generation is fundamental in interpreting regional hydrology and is interconnected with factors that impact humans and the environment. Specifically, flow paths of runoff determine water quality, influence land productivity and are critical in the understanding and prediction of floods (Davie, 2008).

1.1.1 - Identifying and quantifying runoff generation pathways

Runoff generation is dependent on flow pathways that are spatially and temporally variable (Kirkby, 1988), namely: Hortonian or infiltration-excess overland flow (HOF), saturation-excess overland flow (SOF), regolith subsurface flow, and aquifer subsurface flow (Te *et al.*, 1988). For surficial processes, HOF is a dominant runoff pathway when the rate of water input to a given soil exceeds the local infiltration capacity (Horton, 1933; Betson, 1964; Black, 2005), whereas SOF is present when rainfall or snowmelt occurs on saturated soils, i.e., soils with a zero storage capacity (Hewlett and Hibbert, 1967; Dunne and Black, 1970a, 1970b). In hillslope environments or areas defined by artificial drainage, where flow direction is clear, overland flow can be measured using collection troughs (Davie, 2008).

For subsurface processes, specifically regolith subsurface flow and groundwater flow, the main distinction is that the former mainly occurs in the soil vadose zone while the latter takes place in the saturated or phreatic zone only. Regolith subsurface flow includes matrix flow, macropore flow, pipeflow and fill-and-spill flow. These flow types are often described using the generic terms "throughflow" or "interflow" to signify that they are important in feeding water to streams in a transient manner during hydrological events (Weiler, 2005; Davie, 2008). Groundwater flow, because it involves the movement of water in the saturated, phreatic zone, is thought to make up the baseflow component of the hydrograph and feed water to streams during and between hydrological events. Indirect methods to assess regolith subsurface flow and groundwater flow include soil moisture monitoring and water table measurements, while subsurface water levels and subsurface water movement can be directly documented with piezometers (or wells) and nested piezometers respectively, both outfitted with pressure transducers or water level loggers (van der Kamp *et al.*, 2003).

Differentiating surface runoff and subsurface runoff magnitude, frequency, duration and velocity is not straightforward from in-situ measurements alone. Assessing the characteristics of overland flow in areas with un- or poorly defined contributing areas and understanding subsurface water movement is often limited by direct measurement techniques. However, environmental tracers, including stable water isotopes and fluorescent dyes, have been used successfully to quantify and differentiate surface and subsurface runoff generation pathways.

1.1.2 - Factors that influence runoff generation processes

Regional runoff generation pathways are influenced primarily by climate, topography, soils and vegetation. It is often not practicable to study runoff generation pathways and their control over a whole watershed, and hence often the focus is on riparian areas because of their critical importance for runoff and solute transport (McGlynn and McDonnell, 2003; McGlynn and Seibert, 2003). Riparian zones are areas of transition between hillslopes (or floodplains) and streams, i.e., areas where hydrological processes within a watershed converge prior to contributing to streamflow (Swanson *et* *al.*, 1982). As such, assessment of hydrological state variables and control factors in riparian areas can be an effective way to determine whether an active coupling or a decoupling between watershed processes and stream dynamics does exist.

Invariably, runoff generation is a response to hydrological inputs: precipitation quantity dictates input magnitude, while timing, rate and form of precipitation mainly dictate whether water is stored, infiltrated or readily transported as surface or subsurface runoff (Maidment and others, 1992). Landscape topography influences runoff characteristics by determining the direction and rate of surface flow and the location of surface water storage (Davie, 2008). Topography is also important for runoff generation below ground as the slope of the water table is often assumed to be similar to that of surface terrain (e.g., Beven and Kirkby, 1979) . Soil characteristics, soil condition, porosity and structure influence the flow of water through a landscape in general and riparian areas in particular. Soil porosity and structure influence the capacity of a soil to store water and govern infiltration rate and capacity (Brady *et al.*, 1996). Soil condition may change on a seasonal basis: specifically, extended periods of sub-zero temperatures result in frozen soils, which impact runoff generation by reducing the infiltration of water into near-surface soil layers (Zhao and Gray, 1999). Cracks in soils and biopores offer a mechanism for rapid water delivery into lower soil layers (Zhao & Gray, 1999). As for vegetation, it influences infiltration through stemflow, biopores and preferential flow pathways produced by root networks (Davie, 2008).

The importance of spatial patterns of hydrologic state variables – such as near-surface soil moisture – for hydrologic processes at the scale of hydrological events and seasons has also been established (e.g., Grayson *et al.*, 1997; Western *et al.*, 2002; Wilson *et al.*, 2004). For temperate regions of Australia and for rangeland landscapes in particular, Grayson *et al.* (1997) notably suggested that soil water patterns switch between two preferred states that are illustrative of different controls on runoff generation processes. While others have shown the theory of preferred soil moisture states proposed by Grayson *et al.* (1997) to hold true in quasi-pristine forested areas (e.g., James & Roulet, 2007; Ali & Roy, 2010), it is unclear whether preferred states can be discerned for hydrological state variables other than soil moisture (e.g., soil temperature, soil electrical conductivity, frost depth and snow depth) and in environments with different climatic, topographic and soil conditions.

1.1.3 - Runoff generation in the Canadian Prairies

Many interpretations and conceptualizations of runoff generation mechanisms and pathways have been made at the hillslope and small watershed scale based on iconic and mostly forested research sites in the United States such as Coweeta (North Carolina), Sleepers River (Vermont), and Panola Mountain (Georgia). However, the Canadian Prairies do not share similar characteristics with those sites and hence, commonly applied rules linking typical hillslope characteristics with specific dominant runoff generation pathways cannot be applied without modification to the Canadian Prairies. In this region, HOF is traditionally identified as the dominant runoff generation pathway (Fang *et al.*, 2007). The primary hydrological event in the Prairies is the spring snowmelt, which is characterized by rapid surface runoff from snowmelt travelling over frozen ground (Fang *et al.*, 2007). Near-surface soil layers that are commonly frozen during the winter months impede permeability and reduce the capacity for melt-water and early spring rainfall to infiltrate (Brady *et al.*, 1996). The reduced infiltration capacity of frozen soils causes meltwater to flow at the surface, thus leading to HOF generation. The soil moisture content, soil temperature, saturated hydraulic conductivity, soil texture and vegetative cover of an area influences meltwater infiltration into frozen soils (Granger *et al.*, 1984; Zhao and Gray, 1999). Furthermore, snowmelt rate and snow redistribution within the landscape are important in evaluating runoff generation: the release of water stored in snow and ice that exceeds infiltration and surface storage capacities leads to surface runoff (Granger *et al.*, 1984).

Regionally, the depletion of soil moisture from evapotranspiration leaves soils predominantly unsaturated for the majority of the summer (Brady *et al.*, 1996; Fang *et al.*, 2007), making surface runoff generation responses from rainfall events uncommon. High intensity convective storms may produce enough rainfall at a high enough rate to exceed soil infiltration capacity, thus exceptionally resulting in HOF generation (Shook and Pomeroy, 2012). In the Prairies, the distribution of rainfall generated from storms of this nature have a high level of spatial variability (Dyck and Gray, 1979) and typically take place within a single day (Dyck and Gray, 1979; Shook and Pomeroy, 2012). The presence of glacially formed potholes across a large portion of the Prairie landscape, known as the Prairie Pothole Region (PPR), has influenced hydrological studies and conceptualization within the region. A surface fill-and-spill runoff generation mechanism is a key component of hydrology within the PPR. As water volume exceeds pothole storage capacity, excess water spills over and has the potential to travel, as runoff, towards the next pothole along a topographic flow path or towards a neighboring stream.

Frozen ground in general results in a reduction of surface permeability due to soil pores being filled with ice, however different types of frost can have a variable impact on infiltration and percolation (Black, 2005). For instance, concrete frost impacts infiltration when frost reaches the ground surface or when soil above the frost layer is saturated (Striffler et al., 1959) and this can augment natural drainage patterns during the spring snowmelt (Prévost *et al.*, 1990). However, honeycomb and granular frost structures do not necessarily result in decreased soil infiltration capacity, and can actually enhance infiltration (Trimble et al., 1958). In wooded areas, such as those common to areas of south-eastern Manitoba, granular and honeycomb frost has been observed near the surface, while concrete frost appears at greater depths (Striffler et al., 1959). This arrangement of frost types may allow meltwater to infiltrate the soil, eventually reaching a concrete frost impeding layer that may stimulate lateral transmission and runoff (perched subsurface flow). Deformation in near-surface soil layers (e.g., cracks and macropores) interrupts uniformity in soil hydraulic conductivity, creating less restricted pathways for water to move (Black, 2005). Depending on macropore connectivity, meltwater can reach deeper soil layers faster, potentially reaching layers below the frost table and enhancing shallow subsurface water movement (Haupt, 1967; Harris, 1972). Tree cover in areas of the Prairies may result in stemflow; the tree base, root system and surrounding macropores can rapidly deliver water into subsurface soil layers (Davie, 2008). It is worth noting that even in the absence of frost, the literature suggests that the dominant runoff generation mechanism in forested areas is shallow subsurface flow. While flow velocities are usually slow in clay soils, such is not the case in sandy or till-based soils – where hydraulic conductivities are high -, and peat-rich soils that tend to saturate quickly and remain saturated for a long time, making it possible for SOF to occur.

1.2 Research significance and objectives

Runoff generation is fundamental for furthering understanding of Prairie hydrology and has direct influence on seasonal flooding and the delivery of contaminants and nutrients to waterways; factors that ultimately influence regional water management and policy. To date, existing literature for the Canadian Prairies encompasses hydrologic conceptualizations and hypotheses that are centralized on dominant runoff generation mechanisms for the spring snowmelt with primary focus on the Prairie Pothole Region. Absent are assessments of runoff generation in the summer months and eastern Prairies, which may be significantly associated with runoff contributions from shallow subsurface sources. The intended contribution of this M.Sc. thesis is to begin addressing those gaps. The analysis of hydrometric, tracer, and spatial hydrological state variable data from a watershed typical of the astern Prairies is done to test prevailing assumptions of dominant runoff generation mechanisms and potentially suggest that shallow subsurface flow can be a principal contributor to runoff generation in the region.

The objective of this M.Sc. thesis is therefore to examine runoff generation in a Prairie landscape. The primary research objectives examined have, to date, received limited attention in the existing literature, specifically:

1) Determine whether riparian soil moisture, temperature and electrical conductivity can be used to effectively infer vertical and lateral water movements in riparian areas of an eastern Prairie watershed.
2) Assess whether shallow subsurface flow is a significant runoff generation mechanism in an eastern Prairie watershed using a combination of both hydrometric and tracer data.

1.3 Thesis structure

This thesis is separated into two chapters (Chapters 2 and 4) comprising original research presented as independent manuscripts that have been submitted individually for publication. Those chapters describe hydrologic data collection and analysis from the Catfish Creek Watershed, a low relief Prairie watershed situated in south-eastern Manitoba, Canada. Chapter 2 focuses on inferences about riparian-to-stream soil water movement in Prairie landscapes using hydrologic state variables. Chapter 3 is a preliminary synthesis, then followed by the second manuscript, featured in Chapter 4, which focuses on the significance of subsurface flow as a dominant runoff generation mechanism in the eastern Prairies. Lastly, the amalgamated conclusions and synthesis of this thesis in addition to related future opportunities are summarized in Chapter 5.

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CHAPTER 2.

INFERRING VERTICAL AND LATERAL SOIL WATER MOVEMENT IN CANADIAN PRAIRIE RIPARIAN AREAS USING HYDROLOGIC STATE VARIABLES

2.0 Abstract

Flat terrain and soils with variable permeability make it difficult to assess the relative importance of surface and subsurface runoff in the Canadian Prairies, especially in riparian areas that are critical for water transmission and solute transport. The main objective of this study was therefore to determine whether patterns of hydrologic state variables, namely near-surface soil moisture (SM), soil electrical conductivity (SEC) and soil temperature (ST), can help infer riparian-to-stream soil water movement in Prairie landscapes. Focus was on the near-level Catfish Creek Watershed (south-eastern Manitoba, Canada) where three riparian sites were monitored: a naturally vegetated grassland site, a headwater, wooded, forest site and a highly impacted grassed site adjacent to an engineered drainage dyke and a man-made drainage channel. Nine to twelve SM, SEC and ST surveys were completed at each site in 2015 using a 75-point grid. The surveys were then matched with riparian water table data, surface water level data from adjacent drainage channels and indicators of antecedent moisture conditions. Spatial and non-spatial statistics were computed for each state variable pattern and then correlated to indicators of antecedent moisture conditions (AMCs), stream and subsurface water level data to infer soil water movement. Results show that potential evapotranspiration, depth to water table and antecedent precipitation have a significant yet variable impact on SM, SEC and ST patterns. A switching behaviour, between dry and wet conditions, was observed in riparian areas characterized by grassland vegetation and well-drained soils. The occurrence of shallow subsurface flow was inferred during the wettest conditions. While riparian SM conditions were useful for predicting streamflow response in adjacent channels, such was not the case for riparian SEC and ST. Further investigations are however necessary to confirm the usefulness of SM spatial patterns for predicting streamflow response in other landscapes across the Canadian Prairies.

Keywords

Riparian area, soil moisture, soil electrical conductivity, soil temperature, spatial patterns, Canadian Prairies, streamflow response, antecedent moisture conditions

2.1 Introduction

Hydrologic processes are affected by multiple factors or combinations of factors that control their activation, intensity and deactivation (Ambroise, 2004). Because of the difficulties associated with quantifying runoff dynamics and control factor interactions across large scales, several authors have advocated to focus on key, isolated watershed units that are known to regularly collect water and convey it to stream channels (e.g., McGlynn and McDonnell, 2003a,b; McGlynn and Seibert, 2003). In that regard, focusing on near-stream dynamics is important since riparian areas occupy a critical space at the land-water interface within watersheds (Cirmo and McDonnell, 1997; Burt and Pinay, 2005). Near-stream zones are often a hot spot for saturation-excess overland flow, groundwater ridging, translatory flow, macropore flow and other mechanisms known to move significant amounts of both “old” water and “new” water to the stream (Buttle, 1994). As the “last point of contact” before the stream (Hooper *et al.*, 1998), riparian soils have also been shown to modify (i.e., “reset”) the chemical signature of water originating from upslope areas (Robson *et al.*, 1992; Hill, 1993; Peters *et al.*, 1995; Cirmo and McDonnell, 1997). Achieving a better characterization of riparian zones as “chemical modifiers” however requires a good understanding of flow paths and physical mixing processes (e.g.,McGlynn *et al.*, 1999), and the relative importance of surface versus subsurface flowpaths in near-stream zones has been the subject of a long standing debate (e.g., Waddington *et al.*, 1993; McGlynn *et al.*, 1999; McGlynn and McDonnell, 2003a, 2003b). While measuring surface and subsurface flow directly in the field is not straightforward, insights into the activation or deactivation of such pathways can be gained using hydrologic state variables (hereafter referred to as state variables). For instance, soil moisture (SM) has been identified as a state variable that functions both as a factor that influences hydrologic processes (Teuling and Troch, 2005) as well as an indicator of processes that are occurring (Western *et al.*, 1998) and has been used to identify hydrologically active areas (Ambroise, 2004; Huza *et al.*, 2014).

In many published studies, high-density SM measurements have been made across whole watersheds towards pattern analysis and hydrological process interpretation (Ali and Roy, 2010a; Grayson et al., 1997; James and Roulet, 2007; Western et al., 2004, 1998). Spatial patterns of SM have been correlated to antecedent moisture conditions (AMCs) and are considered to be useful indicators of runoff and streamflow response (Grayson *et al.*, 1997; Western *et al.*, 2004; James and Roulet, 2007). The spatial organization of SM has notably been used to put forward a hypothesis suggesting soil water patterns switch between preferred states that are controlled by different landscape characteristics and dominated by different flow processes. The “wet” state occurs when precipitation exceeds evaporation, leading to spatially organized patterns of soil water that reflect a dominance of lateral water movement. Conversely, the “dry” state occurs when evapotranspiration exceeds precipitation and spatially disorganized (or random) soil water patterns indicate vertical water movement. The absence of significant lateral flow in the “dry” state also infers hydrologic disconnectivity between a point in the watershed and its upslope area. The dominance of either of the preferred states is dependent on the balance between precipitation and evapotranspiration, while the switching behavior between states is dependent on both this balance and the soil water storage deficit (Grayson *et al.*, 1997). It has been noted that some environments may be totally dominated by one of the states without exhibiting switching behavior (Grayson *et al.*, 1997).

While the preferred states hypothesis is an interesting framework for the conceptualization of subsurface flow processes, its universal applicability is unclear. The extent to which spatial patterns acquired in specific watershed units – such as riparian areas – rather than whole watersheds are sensitive enough to allow the application of that framework remains unknown. Furthermore, some of the principles underlying the preferred states hypothesis have been challenged as a result of contrasting evidence regarding SM dynamics. Some studies have suggested that spatial variability in SM increases with decreased SM content, while others have rather shown that spatial variability increases with increased SM content (Teuling and Troch, 2005). Some authors also believe that SM patterns are representative of lateral subsurface flow during events (Grayson *et al.*, 1997) while others have rather argued that SM development in time and space may not be representative of water movement (van Meerveld and McDonnell, 2005). It is also worth noting that while the importance of SM patterns for process interpretations has been extensively discussed in the hydrologic literature (Grayson *et al.*, 1997; Western *et al.*, 2004; van Meerveld and McDonnell, 2005; Huza *et al.*, 2014), less attention has been dedicated to other potentially important state variables such as soil temperature (ST) and soil electrical conductivity (SEC). ST is known to influence meltwater infiltration into frozen soils (Dyck and Gray, 1979; Granger *et al.*, 1984; Zhao and Gray, 1999) and dictates the type of soil frost (e.g., concrete, honeycomb and granular) that develops during the winter months (Striffler, 1959). ST is also a key variable in determining the land surface heat and water balance, evapotranspiration and changes in SM in both winter and non-winter months (Lakshmi *et al.*, 2003). Similarly, SEC has been used as a proxy for soil water content and soil texture (Corwin and Lesch, 2005; Davie, 2008) and it can be a useful indicator of processes resulting in elevated soil salinity, such as groundwater exfiltration (Todd *et al.*, 1980) and evapoconcentration (Tanji, 2002). The SM-focused preferred states hypothesis was initially suggested for quasi-pristine temperate regions with rolling hills (Grayson *et al.*, 1997) and it has since been verified in other forested areas (e.g., Ali and Roy, 2010a). However, it has yet to be tested not only in areas with different landscape and climatic conditions but also across different spatial scales or with different state variables.

The Canadian Prairie Region (hereafter referred to simply as the Prairies) are an example of a landscape not only with peculiar characteristics that differ from landscapes where the preferred states hypothesis was previously tested, but also where riparian areas are of particular importance. Land-use conversion as well as widespread stream and lake eutrophication in the Prairies have led to increased interest in riparian area hydrology (Dodds and Oakes, 2006; Schilling and Spooner, 2006; Banner *et al.*, 2009) – based on the assumed ability of near-stream zones to capture nutrients and buffer stream chemistry (Burt and Pinay, 2005; Dosskey *et al.*, 2010). The Prairies cover an extremely large portion of south central Canada and the north central United States and are characterized by a near-level topographic profile (Welsted *et al.*, 1996) and significant human modification to the landscape, including engineered drainage channels and roads. In this environment, snow acts as a form of water storage and is vital for SM recharge as annual SM deficits are the norm (Shook and Pomeroy, 2012). Seasonally frozen near-surface soil layers reduce meltwater infiltration (Brady *et al.*, 1996), and snowmelt produces approximately 80% of the total annual runoff (Gray and Landine, 1988). High infiltration capacity and limited rainfall results in little surface runoff generation in the summer months; only the most intense rainfall events result in Hortonian overland flow (Gray, 1973), and exceptionally wet conditions can be conducive to shallow subsurface flow and saturation-excess overland flow generation. One characteristic of particular importance is that the flat and hummocky terrain typical of the Prairies makes it extremely rare for headwaters to be hydrologically connected to watershed outlets (Godwin and Martin, 1975; PFRA-Hydrology Division, 1983; Martin, 2001). Event-specific effective drainage areas – although difficult to quantify – are known to be much smaller than gross drainage areas (defined by surface topography once all small depressions associated with hummocky terrain have been filled within a digital elevation model). In such Prairie environments, the only area where interactions between landscape characteristics, runoff generation processes and streamflow response are the most evident is at the land-water interface, namely the riparian area. Hence, while most watersheds around the World are characterized by hydrographs which are a convolution of hillslope and riparian contributions to the stream (e.g., McGlynn and McDonnell, 2003a, 2003b; McGlynn *et al.*, 2004), the quasi-permanent disconnectivity of headwaters in the Prairies has reinforced the idea that efficient flood and water quality mitigation must be achieved through the management of riparian areas (Sheppard *et al.*, 2006; Alan Stewart and Agriculture and Agri-Food Canada, 2013). The aim of the current chapter was therefore to use hydrologic state variables to infer vertical and lateral water movements in riparian areas in a non-typical landscape where macro-scale topography is not assumed to be the main driver of runoff generation. The focus was on typical eastern Prairie riparian areas with different meso-topography and vegetation as well as various levels of human alteration. Riparian area monitoring was done across a range of climatic conditions to ensure that different degrees of riparian-stream connectivity could be captured. Three objectives were pursued:

1) Characterize the spatial and temporal variability in riparian soil moisture (SM), soil electrical conductivity (SEC) and soil temperature (ST).
2) Assess the influence of antecedent moisture conditions on the statistical distribution and spatial organization of riparian SM, SEC and ST.
3) Determine whether riparian SM, SEC and ST pattern characteristics are good predictors of streamflow response.

2.2 Methods

2.2.1 - Study site description

The 642 km^2 Catfish Creek Watershed (CCW) is a meso-scale watershed located approximately 100 km north east of the city of Winnipeg (Manitoba, Canada) (Figure 2.1). The region is characterized by hummocky, crystalline Archean bedrock which is covered by a discontinuous mantle of sandy glacial till veneer with morainal uplands as well as extensive fen peatlands (Smith *et al.*, 1998). The natural condition of the CCW has been altered through wetland drainage for agricultural expansion and the addition of road and engineered drainage networks (Elliott, 1977). The watershed has a near-level topographic profile (maximum relief of 91.4 m, average slope of 1.4 degrees or 2.5%) and land use is equally split between agricultural and forested land. In terms of climate, the CCW has short warm summers and long cold winters with a mean annual temperature of 1.9 °C and an average growing season length of 180 days (Environment Canada, 2011). Mean annual precipitation is roughly 530 mm with substantial variability from year to year and increased precipitation near the watershed outlet and Lake Winnipeg (Environment Canada, 2011). Approximately one fifth of annual precipitation falls as snow within the region and average annual moisture deficits are around 90 mm. On a geologic time scale, natural waterways in the CCW are relatively young and have been incised in deposits left by glacial activity and the drainage of Glacial Lake Agassiz (Teller and Leverington, 2004). The brief history of these waterways plays a significant role in the limited hydrologic connectivity in the region (Bloom, 1998).

Three riparian areas – selected based on their typical albeit contrasting characteristics within the Prairie context – were instrumented and monitored in the CCW (Figure 2.1). Here the goal was to look solely at the riparian scale where water-driven connectivity to the stream is more frequently expressed and also easier to assess using spatially detailed field measurements. However, the intent was not to use the riparian-scale results as a way to upscale or generalize conclusions for the whole CCW. Delineation of the targeted riparian areas was not done based on topographic changes in the land-stream continuum – which are relatively subtle – but rather based on vegetation changes which were assumed to reflect either changes in subsurface water and nutrient availability for plant growth or correspond to the human-imposed boundaries between "natural" riparian areas and cultivated crops. Site A is naturally vegetated (mixed grass species), has an average slope between 5-9% and is considered to be well drained with Solonetzic Gray Luvisol soils developed on moderately to strongly calcareous lacustrine clay (GIS4AG, 2015; Soil Science Society of America, 2016) adjacent to a man-made drainage channel. Site B is a naturally forested site adjacent to a small headwater creek and located in the eastern region of the watershed near the interface of the Boreal Prairies and Boreal Shield ecozones (Smith *et al.*, 1998). Site B has an average slope between 5-9% and has poorly drained Rego Humic Gleysol soils developed on strongly calcareous, stony glacial till (GIS4AG, 2015). As for Site C, it is located in the southern portion of the watershed in a lowland where agriculture is the main land use. Site C is adjacent to an engineered drainage dyke and a man-made drainage channel: it has an average slope between 0-2% and is considered to be poorly drained with Terric Messisol soils developed on mesic fen peat overlying calcareous loam to clayey lacustrine sediments (GIS4AG, 2015).

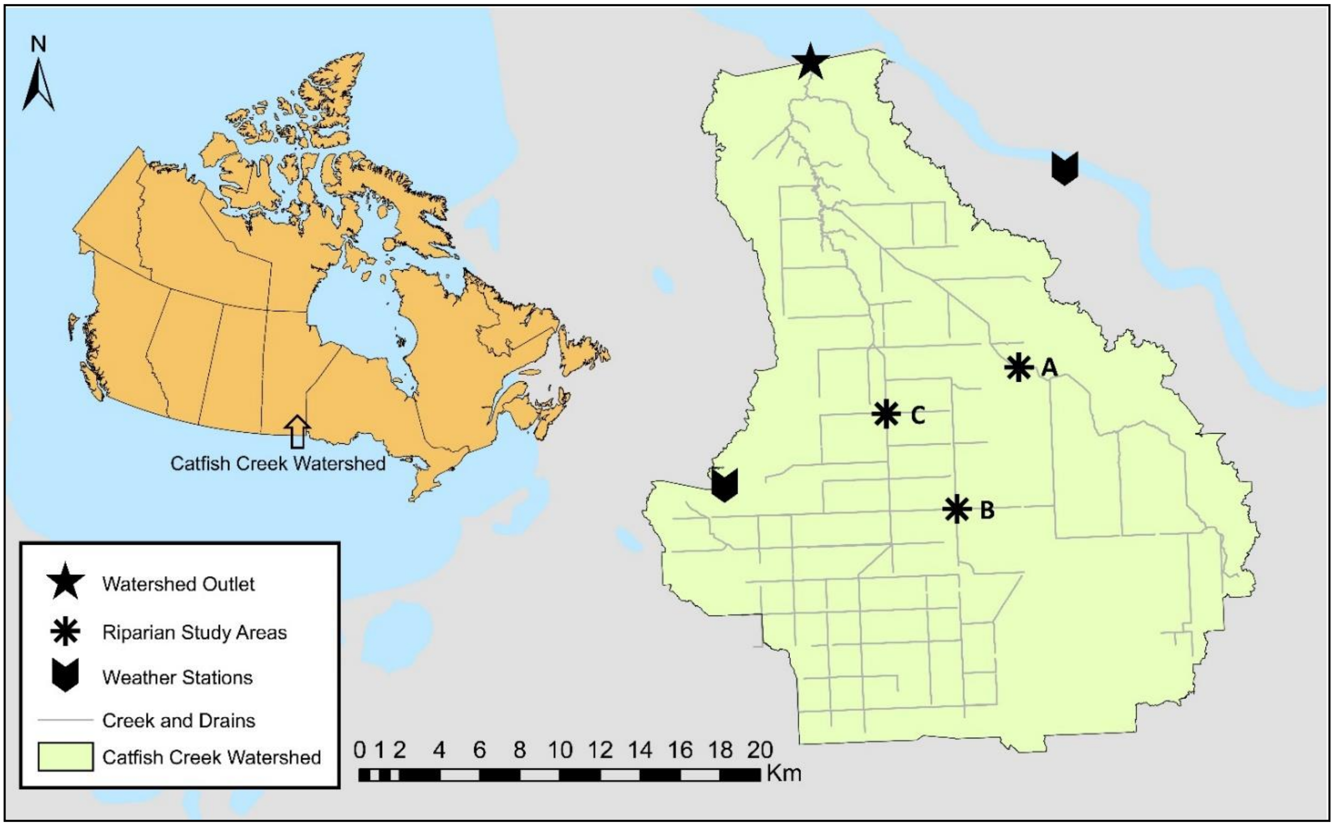

Figure 2.1. Location of the Catfish Creek Watershed as well as the riparian areas and weather stations used in this study.

2.2.2 - Experimental setup and data collection

All three riparian area sites shared the same experimental setup focused on a 5 m x 15 m rectangular plot oriented perpendicularly to the adjacent channel (Figure 2.2). A sampling grid with 75 equally spaced nodes was established in each rectangular plot for manual surveys of target variables (see below). Each plot was outfitted with three groundwater wells (1.5 m deep) on the upstream edge and a stilling well in the drainage channel downstream from each plot.

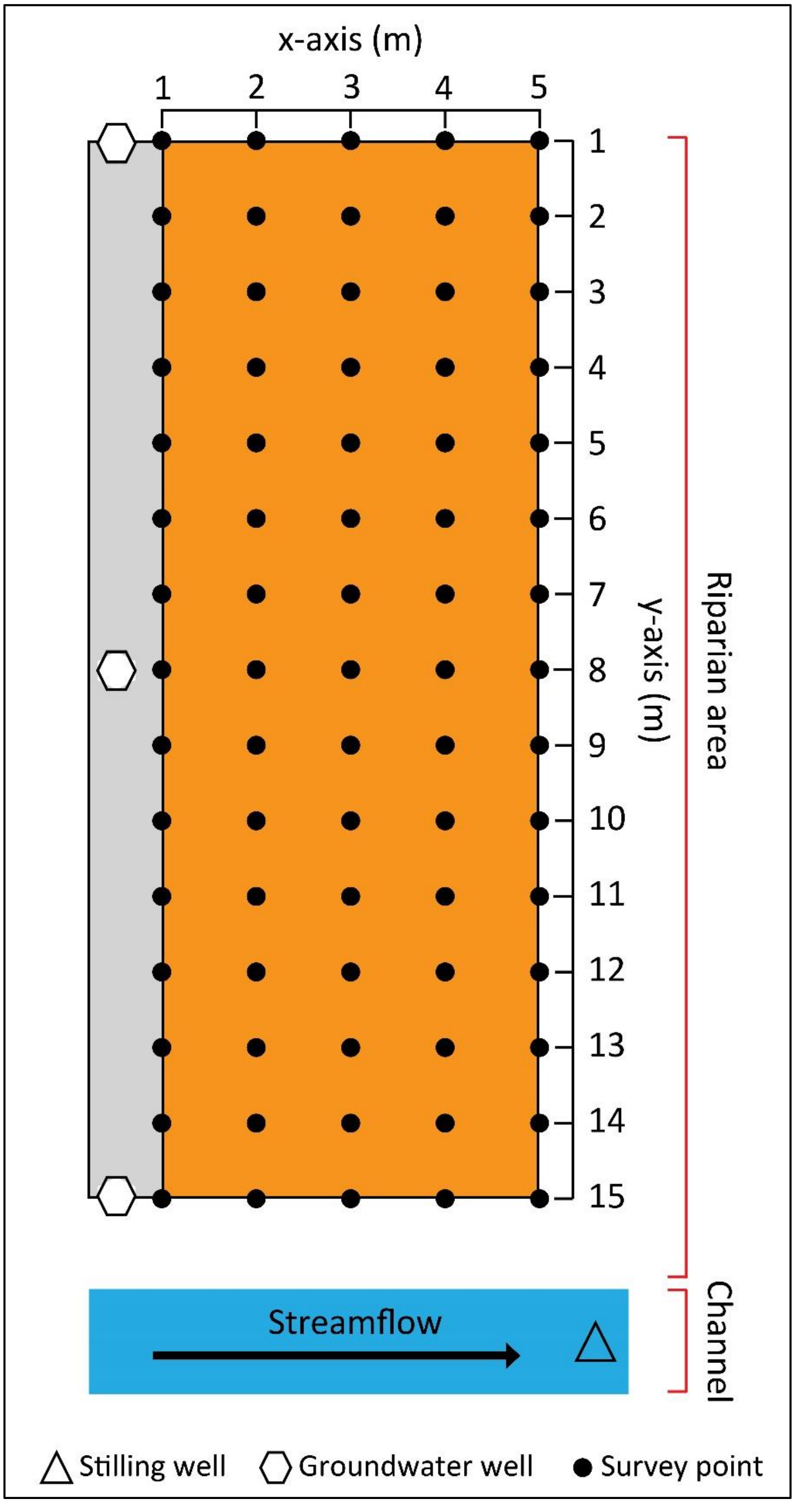

Figure 2.2. Experimental setup used for the three studied riparian areas. All groundwater wells are 1.5 m deep. Groundwater wells and stilling wells were outfitted with capacitance-based water level loggers.

Prior to SM, SEC and ST data collection, elevation models of the three riparian sites were created using a Sokkia Total Station Set 4110 and used to evaluate potential meso-topographic influences on SM, SEC and ST. Leading up to the 2015 spring snowmelt, spatial surveys of depth to the frost table (FD) and snow depth (SD) were completed at each site on nine and seven occasions respectively (Table 2.1). SD was simply measured from the ground surface to the top of the snowpack surface using a measuring stick. FD was measured using a pick that was forced vertically though the soil profile until refusal (presumably caused by a confining layer of ground frost). Both FD and SD were recorded at each of the 75 grid nodes. SWE was also measured at each site on three occasions prior to the 2015 snowmelt but only at the four corners of the experimental plot to avoid coring (and hence disturbing) the snowpack in a major way before the onset of snowmelt. Site-specific averages of snow density were then derived from the SWE measurements and used to convert the 75-point SD surveys to SWE maps. Spatial surveys of SM, SEC and ST were completed for all sites at a depth of 30 cm using a capacitance-based probe (AQUATERR Instruments & Automation). At sites A and B, those surveys were completed once in fall 2014 – to characterize riparian conditions before winter freeze-up – and on 12 occasions in spring and summer 2015 (Table 2.1). Surveys were completed at Site C in a similar manner, with the exception of three surveys in early spring that were discarded or not completed due to vandalism close to the study area (Table 2.1). Volumetric moisture content in the top 30 cm of the soil profile was measured with possible values ranging from 0 (dry soils) to 100% (saturated soils). SEC and ST were measured in μS/cm and °C, respectively.

Table 2.1. Summary of survey dates and variables mapped at each of the three riparian study areas.

Survey Date	Site A	Site B	Site C
11/03/14	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
03/08/15	SD, FD, SWE	SD, FD, SWE	SD, FD, SWE
03/11/15	SD, FD, SWE	SD, FD, SWE	SD, FD, SWE
03/13/15	SD, FD, SWE	SD, FD, SWE	SD, FD, SWE
03/16/15	SD, FD	SD, FD	SD, FD
03/20/15	SD, FD	SD, FD	SD, FD
03/27/15	SD, FD	SD, FD	SD, FD
04/15/15	SD, FD	SD, FD	SD, FD
05/06/15	SM, SEC, ST, SD, FD	SM, SEC, ST, SD, FD	No data
05/08/15	SM, SEC, ST, SD, FD	SM, SEC, ST, SD, FD	No data
05/11/15	SM, SEC, ST	SM, SEC, ST	No data
05/13/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
05/15/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
05/19/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
05/22/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
05/27/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
06/30/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
07/23/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
07/26/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST
07/28/15	SM, SEC, ST	SM, SEC, ST	SM, SEC, ST

In parallel to synoptic survey data for SM, SEC, ST, SD and FD, water level data was recorded at a 15-minute frequency in spring and summer 2015 using Odyssey^{TM} capacitance-based water level loggers: the loggers deployed in groundwater wells were used to record the depth to the water table (DWT) while those deployed in stilling wells were used to measure stream water levels (SWL) which were then converted to discharge (Q) using site-specific rating curves. Site-specific surrogate measures of antecedent moisture conditions (AMCs) and indicators of streamflow response were estimated for each of the survey dates (Table 2.2). AMC measures were based on temperature and precipitation data from two nearby weather stations (Figure 2.1) as well as local DWT data. Potential evapotranspiration (PET) was estimated for each survey date and location using the Hamon formula (Hamon, 1961), and DWT-related and rainfall-related measures were computed over a range of antecedent temporal windows (e.g., x days before each survey, with x varying between 1 and 30) in order to capture short- (1-3 days), medium- (5-10 days) and long-term (14-30 days) meteorological influences on hydrologic state variables. For each site, inverse distance weighing (Price *et al.*, 2000) was applied to meteorological variables to account for any effect due the relative distance of each site from the two weather stations.

Table 2.2. Surrogate measures of antecedent moisture conditions and indicators of streamflow response computed in relation to each survey at all three sites. "Temporal window" refers to the antecedent temporal window used for each computation.

Surrogate Measures of Antecedent Moisture Conditions									
	Description	Temporal window (Days)							
		0	1	2	3	5	7	14	30
AAT	Average air temperature on day of survey	X							
PET	Potential evapotranspiration on day of survey	X							
AP_{xday}	Antecedent precipitation for fixed period (x days) prior to survey		X	X	X	X	X	X	X
DWT_{mean}	Average depth to water table on day of survey	X							
$DWT\text{-}30_{xday}$	% of time water table is within 30 cm of surface for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$DWT\text{-}100_{xday}$	% of time water table exceeded surface for fixed period (x days) prior to survey		X	X	X	X	X	X	X
Indicators of Streamflow Response									
	Description	Temporal window (Days)							
		0	1	2	3	5	7	14	30
SWL_{mean}	Average stream water level on day of survey	X							
Q_{mean}	Average stream discharge on day of survey	X							
$SWL\text{-}0_{xday}$	% of time stream channel was dry for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$SWL\text{-}25_{xday}$	% of time stream channel was more than 25% full for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$SWL\text{-}50_{xday}$	% of time stream channel was more than 50% full for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$SWL\text{-}75_{xday}$	% of time stream channel was more than 75% full for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$SWL\text{-}100_{xday}$	% of time stream channel was more than 100% full for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$SWL\text{-}R_{xday}$	% of time stream water level was rising for fixed period (x days) prior to survey		X	X	X	X	X	X	X
$SWL\text{-}F_{xday}$	% of time stream water level was falling for fixed period (x days) prior to survey		X	X	X	X	X	X	X

2.2.3 - Data processing

To characterize the spatial and temporal variability in SM, SEC and ST (objective 1), both non-spatial and spatial statistics were used. The statistical variability of site-specific and survey-specific

SM, SEC and ST was assessed using measures of central tendency (MCT, i.e., mean and median of each variable), and measures of dispersion (MDS, i.e., minimum, maximum and standard deviation of each variable). These statistics, although non spatial, have been used previously with SM maps in particular to quantify the amount of intra-pattern variability (Western *et al.*, 2004). In addition to the non-spatial statistics, the geostatistical correlation structure of site-specific and survey-specific SM, SEC and ST patterns was also estimated. Experimental variograms, that express the data variance (γ) as a function of separation distance or lag (h) in a given direction (Moulin, 2003), were built:

$$\gamma\,(h) = \frac{\sum(y(x) - y(x+h))^2}{2N}$$

where N is the number of pairs of experimental observations; y(x) is the y property value at location x; and y(x + h) is the y property value at location separated by a distance h from location x. Variograms have been used to describe the variance between measures of SM at two different points in a spatial field as a function of their separation and to assess whether differences in geostatistical structure of SM patterns can be used to detect differences in hydrologic processes (Western *et al.*, 1998, 2004; Ali and Roy, 2010a). Similarly, relations between other state variables (e.g., SEC and ST) and physical soil properties have been examined using variograms (Davidoff *et al.*, 1986; McBratney and Webster, 1986; Carroll and Oliver, 2005). At each site and for each survey, the spatial autocorrelation of SM, SEC and ST data was assessed in all directions as well as perpendicularly to the stream channel using the range (i.e., average distance between points after which autocorrelation no longer exists (Royle, 1980)) of omnidirectional and directional variograms. To obtain the range, a variogram model was fitted to the survey data through visual inspection. In most cases, the exponential, Gaussian and linear models led to the best fit; however, a variogram range could only be computed for data that was represented using the exponential and Gaussian models.

To assess the influence of AMCs on the statistical distribution and spatial organization of SM, SEC and ST (objective 2), and to determine whether SM, SEC and ST pattern characteristics are good predictors of streamflow response (objective 3), various correlation analyses were carried. First, the potentially nonlinear association between surrogate measures of AMCs (Table 2.2) and both spatial and non-spatial statistics of SM, SEC and ST (see above) was examined by computing Spearman rank correlation coefficients (ρ) and associated p-values between pairs of variables. The Spearman coefficient was used not only because it is suitable to quantify both linear and nonlinear relations between pairs of variables but also because it does not assume normal distribution of the data (Sokal and Rohlf, 1995). In relation to objective 2, the current study also aimed at quantifying how soil condition factors such as FD and SWE, which are known to vary over fine spatial scales, might influence patterns of SM, SEC and ST. For the first five surveys of SM, SEC and ST following the onset of snowmelt, the spatial interaction between indicators of soil condition (i.e., SWE and FD patterns) and the three targeted state variables (i.e., SM, SEC and ST) was assessed using cross-correlation analysis. A cross-correlation length was computed as the lag distance between two points for which the correlation between indicators of soil condition and the hydrologic state variable was at its maximum. The correlation coefficient associated with that distance was also recorded. Lastly, the potential for using state variables to predict streamflow response was evaluated by computing Spearman rank correlation coefficients between spatial and non-spatial pattern statistics and indicators of streamflow response (Table 2.2). With the exception of geostatistical analyses performed in Surfer (Version 12), all statistical analyses were performed in MATLAB (R2015).

2.3 Results

2.3.1 - Variability in meso-topography, water level dynamics and state variables across sites

Low-relief terrain was observed at all three riparian areas (Figure 2.3), with the dominant slope direction perpendicular to the adjacent channels. Overall, Sites A and C had rather smooth topographic surfaces, with elevation gradually decreasing towards the channel. A flat plateau along the edge of the study area farthest from the channel was also present at Site A. The meso-topographic surface of Site B was rather irregular with more pronounced meso-topographic depressions and an elevated planar mid-section (Figure 2.3). For Site A, timeseries of SWL and DWT showed synchronicity between stream and subsurface dynamics, although small-scale fluctuations were less visible in the DWT timeseries (Figure 2.4). SWL and DWT small-scale (or short-term) fluctuations were not always synchronous at Sites B and C (Figures 2.5 and 2.6).

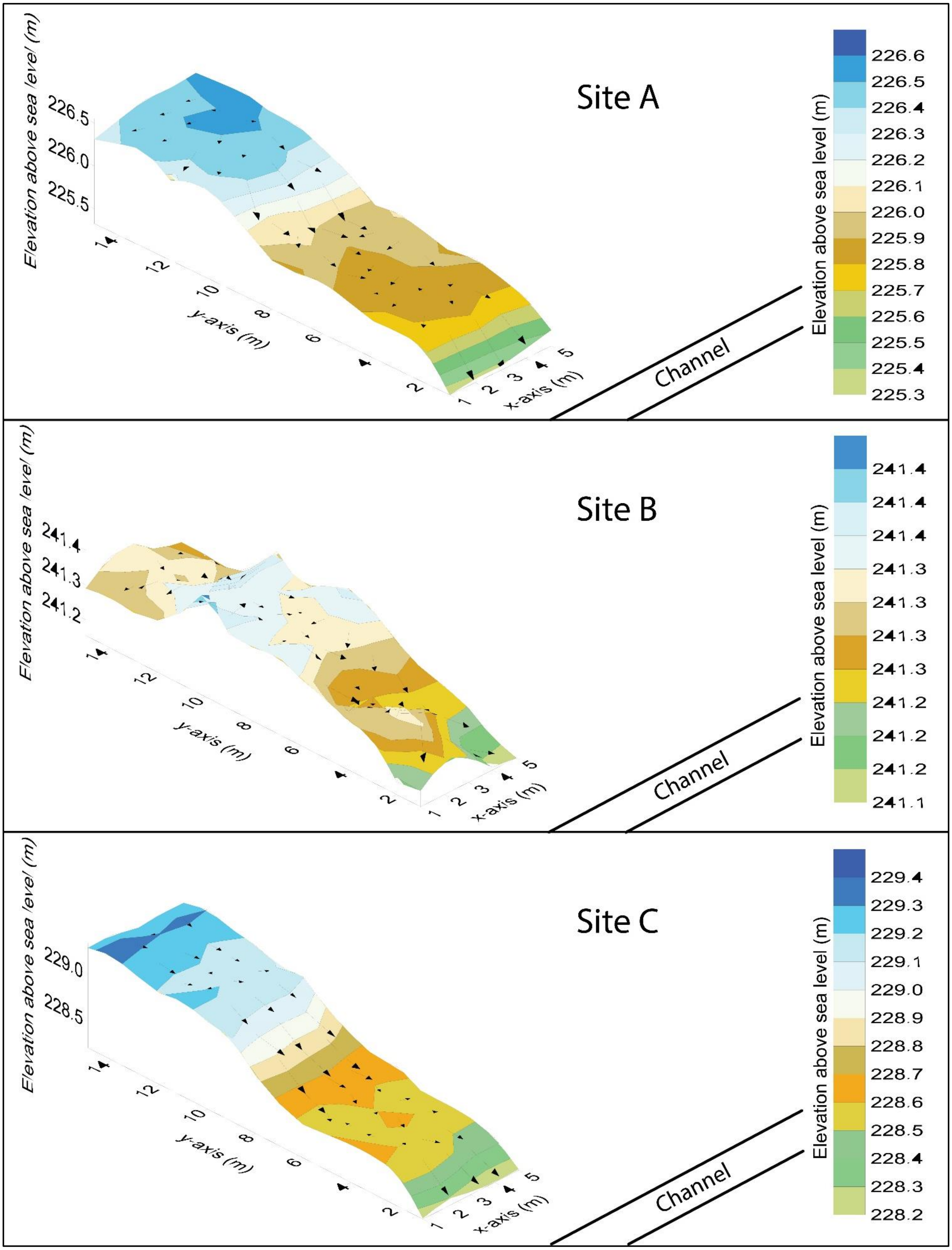

Figure 2.3. Meso-topography associated with the three studied riparian areas. Small black arrows indicate the direction and magnitude of slope within each survey area

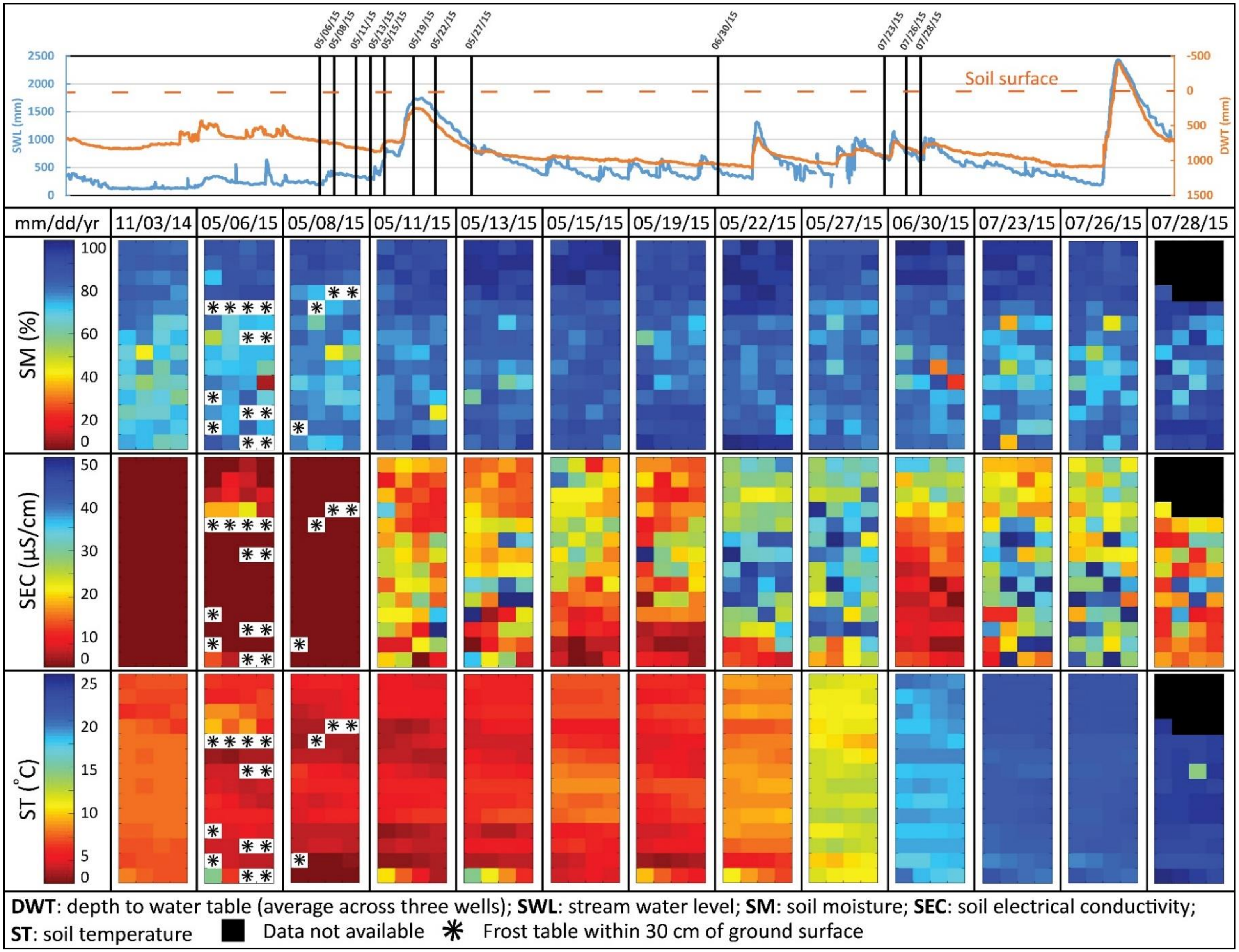

Figure 2.4. Soil moisture, soil electrical conductivity and soil temperature maps associated with surveys completed at Site A. For each map, the orientation of the experimental plot is the same as shown in Figure 2.2. The overlaying timeseries show adjacent stream water level and depth to the water table for Site A for the 2015 open water season.

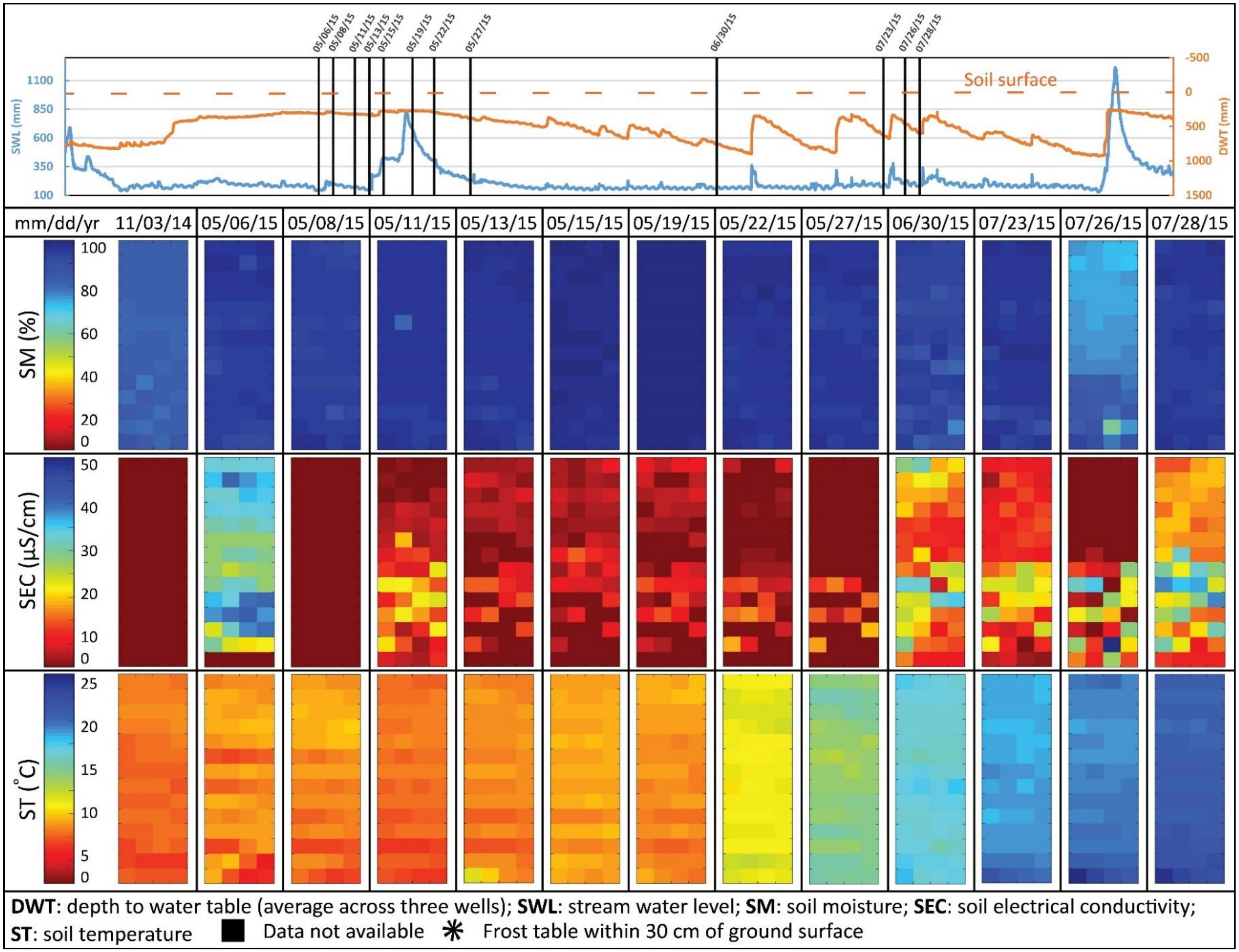

Figure 2.5. Soil moisture, soil electrical conductivity and soil temperature maps associated with surveys completed at Site B. For each map, the orientation of the experimental plot is the same as shown in Figure 2.2. The overlaying timeseries show adjacent stream water level and depth to the water table for Site B for the 2015 open water season.

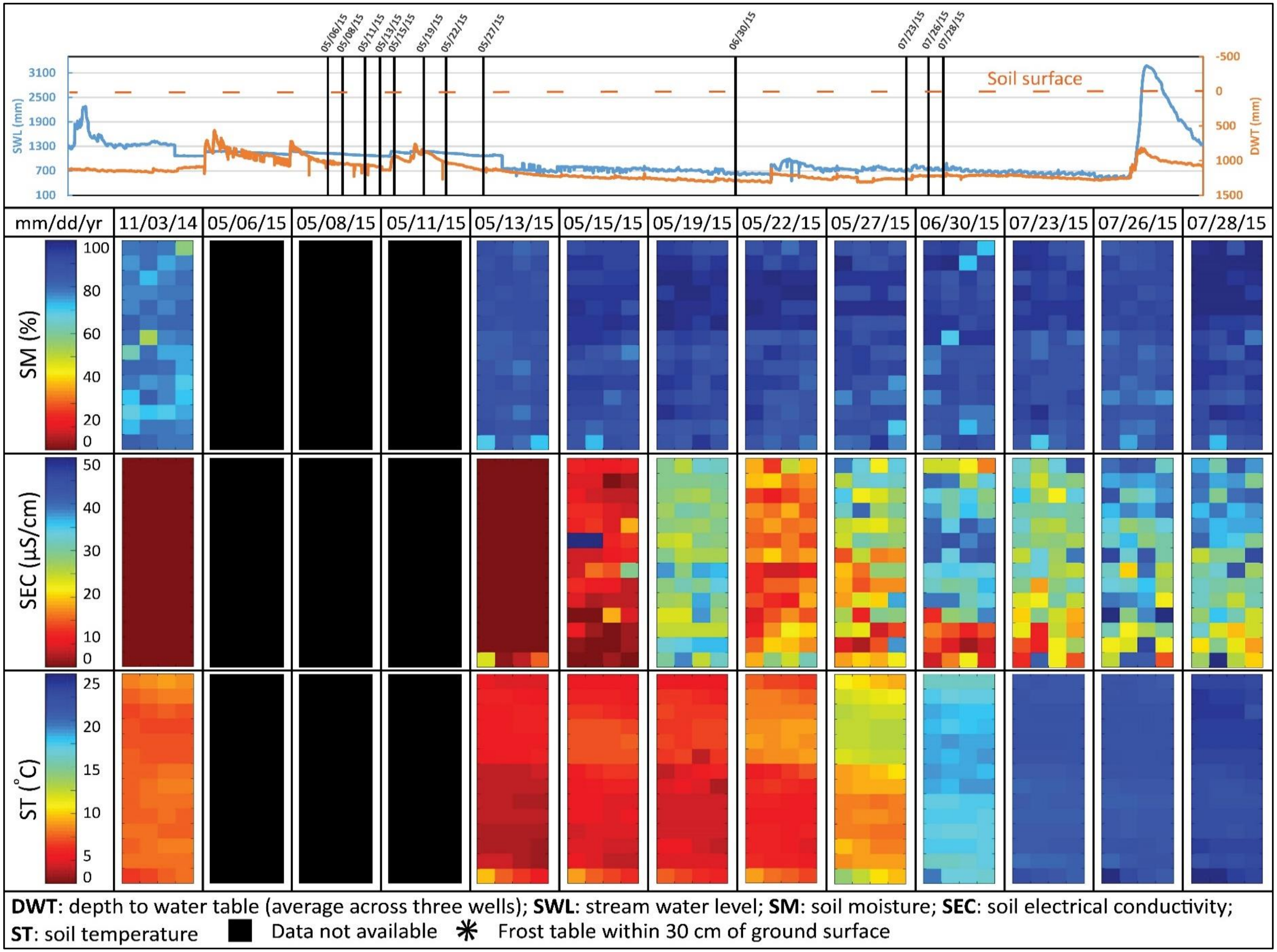

Figure 2.6. Soil moisture, soil electrical conductivity and soil temperature maps associated with surveys completed at Site C. For each map, the orientation of the experimental plot is the same as shown in Figure 2.2. The overlaying timeseries show adjacent stream water level and depth to the water table for Site C for the 2015 open water season.

At Sites A and C (Figures 2.4 and 2.6), heterogeneous SM conditions with two distinct regions were observed for most surveys: a wetter region on the far edge of the riparian area and a drier region with a more random SM distribution closer to the drainage channel. The heterogeneous patterns at Sites A and C were strongly contrasted with the continuously and homogeneously wet patterns observed at Site B (Figure 2.5). During mid-summer at Sites A and C, a distinct switch from heterogeneous SM conditions to wetter, more homogeneous conditions was observed and coincided with smaller DWT values and a peak in SWL (Figure 2.4). Site B was associated with a small SM standard deviation and SM range (Figures 2.7), reflecting its homogeneously wet patterns, while Sites A and C had larger range and standard deviation values, indicating greater variability (higher heterogeneity) in survey-specific SM values. The spatial organization of SEC at Sites A and C (Figures 2.4 and 2.5) seemed random, as opposed to Site B (Figure 2.5) that showed a structured SEC distribution with an area of elevated SEC at the center of the riparian area and lower SEC elsewhere. Non-spatial statistics (Figure 2.7) showed higher SEC values at Sites A and C with greater pattern variability compared to Site B. For all sites, ST patterns were typically well organized with a clear increase in soil temperatures from spring to summer (Figures 2.4, 2.5 and 2.6). At Site B, ST patterns were relatively homogeneous within the survey area, while at Sites A and C, ST increased with distance from the stream during the spring and became more spatially homogeneous during the summer.

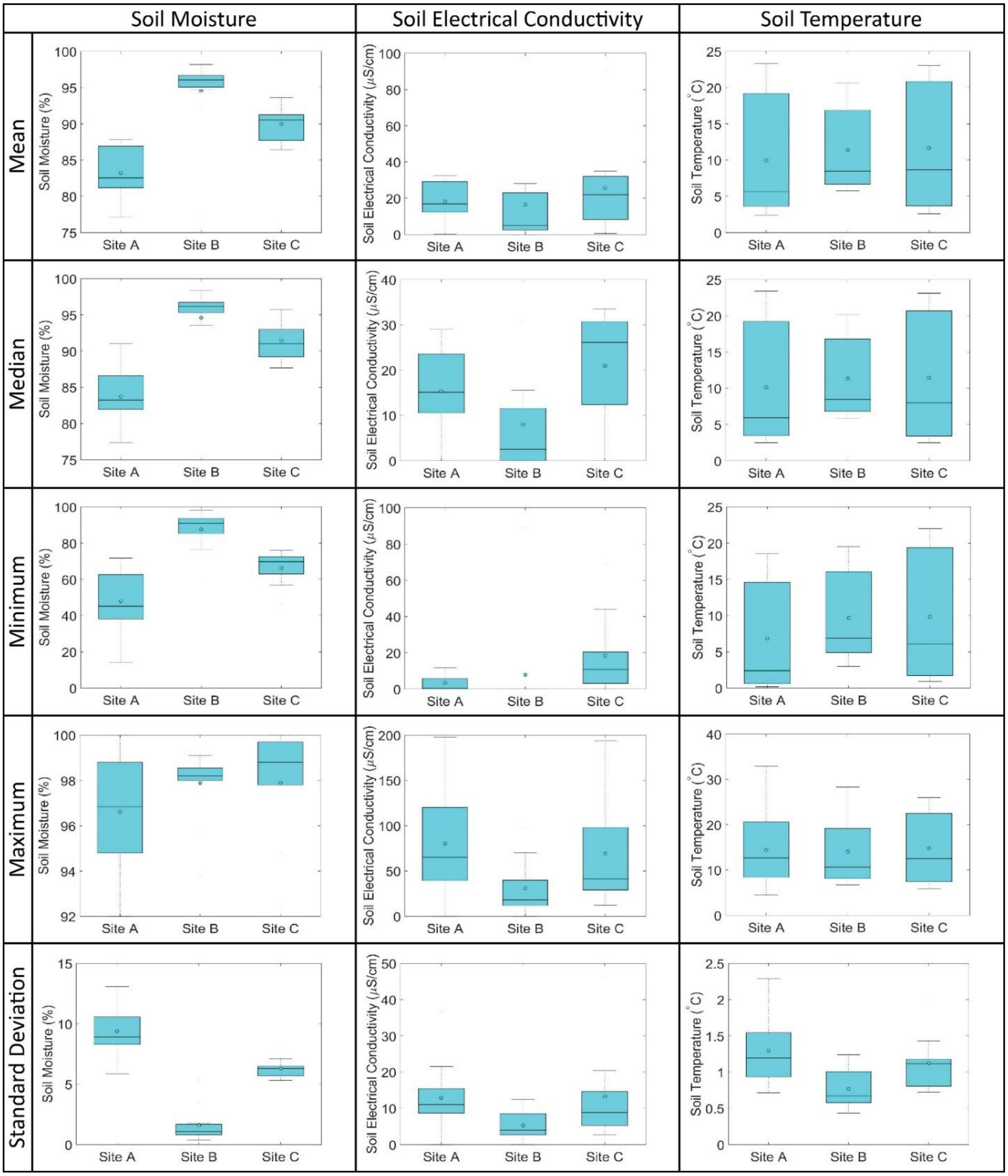

Figure 2.7. Box-and-whisker plots summarizing the variation (over time) of pattern statistics, i.e., mean, median, minimum, maximum and standard deviation of soil moisture, electrical conductivity and temperature. Each box has lines at the lower quartile, median, and upper quartile values, while the whiskers show the extent of the remaining data (minimum and maximum). Circles within each box represent the sample mean.

Larger median directional variogram ranges were found, compared to omnidirectional ranges for SM and SEC at both Sites A and C (Figure 2.8). For ST, site-specific omnidirectional ranges were rather similar to directional ranges. For all state variables, greater inter-survey variability in omnidirectional variogram ranges, compared to directional variogram ranges, was observed at Sites A and C.

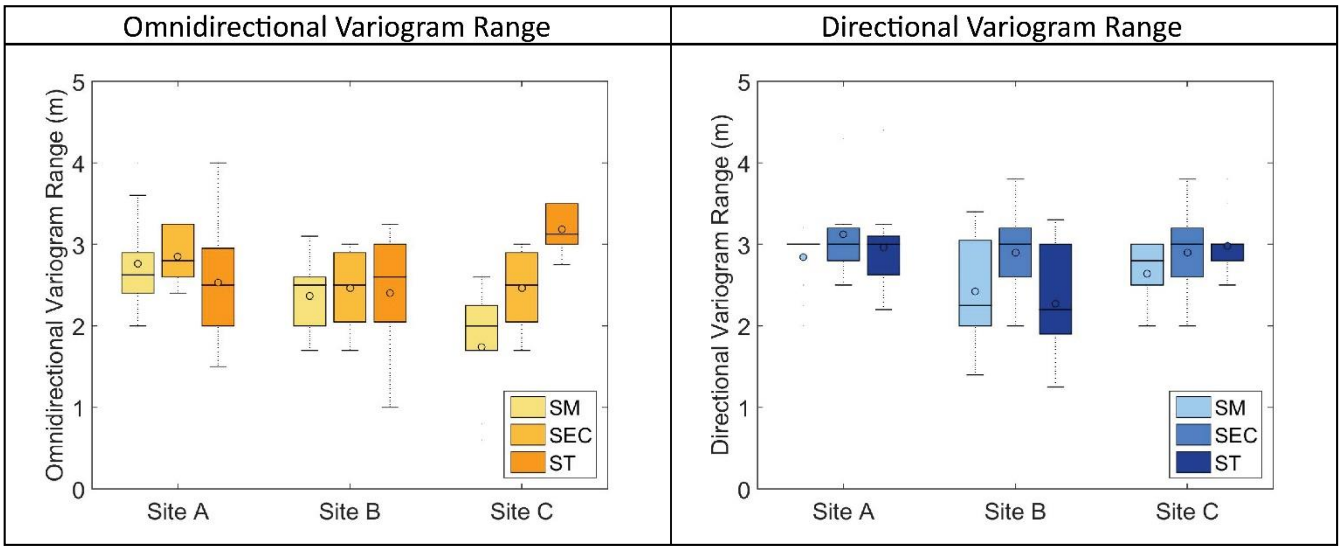

Figure 2.8. Box-and-whisker plots showing the variation (over time) of the omnidirectional and directional variogram ranges for soil moisture, soil electrical conductivity and soil temperature at each site. Each box has lines at the lower quartile, median, and upper quartile values, while the whiskers show the extent of the remaining data (minimum and maximum). Circles within each box represent the sample mean.

2.3.2 - Influence of AMCs and indicators of soil condition on state variables

At Sites A and C, non-spatial statistics of state variables (Table 2.3) were correlated with some measures of antecedent precipitation (Figure 2.9); only correlations significant at the 95% level were kept for interpretation. For Site A, non-spatial SM statistics (MCT) were correlated with AP_5, AP_7 and AP_{14} (max $\rho = 0.66$); mean SEC was correlated with AP_{30} only ($\rho = 0.56$); while non-spatial ST statistics (MCT measures) were correlated with AP_{10}, AP_{14} and AP_{30} (max $\rho = 0.79$). Similarly, for Site C, non-spatial SM statistics (MCT measures) were correlated with AP_7 and AP_{30} (max $\rho = 0.66$); and non-spatial SEC statistics (MCT measures and minimum) were correlated with all antecedent precipitation values except AP_1 and AP_2 (max $\rho = 0.90$). There was also a statistically significant and positive relation between PET and MCT measures for both SEC and ST at Sites A and C (max $\rho = 0.82$ and 0.86 respectively). Conversely, at Site B, non-spatial SM and SEC statistics were correlated with DWT_{mean} (max $\rho = 0.69$) and PET (max $\rho = 0.84$) respectively, while ST was not correlated with any surrogate measure of AMCs. Analysis of the relation between survey-specific spatial statistics (omnidirectional and directional variogram ranges) and DWT revealed a statistically significant positive correlation ($\rho = 0.57$) between DWT_{mean} and directional variogram ranges for SEC at Site C. Also, significant relations between directional variogram ranges for SEC and a small number of AP indicators were observed at Sites A and B (max $\rho = 0.85$).

Table 2.3. Summary of Spearman rank correlation analysis conducted between soil moisture, soil electrical conductivity and soil temperature pattern characteristics, surrogate measures of antecedent moisture conditions and indicators of streamflow response. Descriptions of abbreviations can be found in Table 2.2. X: statistically significant positive correlation (95% level); O: statistically significant negative correlation (95% level). Stdev: standard deviation.

		Mean			Median			Minimum			Maximum			Stdev		
		SM	**SEC**	**ST**	**SM**	**SEC**	**ST**	**SM**	**SEC**	**ST**	**SM**	**SEC**	**ST**	**SM**	**SEC**	**ST**
Site A	PET		X	X			X			X			X			
	AP_{1day}															
	AP_{7day}	X			X											
	AP_{30day}		X	X	X		X			X			X			
	Q_{mean}							X								
	SWL_{mean}							X								
	SWL-25_{7day}		X	X		X	X	X	X	X						
	SWL-50_{7day}		X	X		X	X	X	X	X						
	SWL-75_{7day}					X		X	X							
	SWL-100_{7day}							X						O		
	SWL-R_{7day}															
	SWL-F_{7day}			X						X			X			
	DWT_{mean}															
	DWT-30_{7day}							X						O		
Site B	PET		X						X			X			X	
	AP_{1day}															
	AP_{7day}															
	AP_{30day}															
	Q_{mean}	X						X			X			O		
	SWL_{mean}	X						X			X			O		
	SWL-25_{7day}															
	SWL-50_{7day}							X			X			O		
	SWL-75_{7day}										X					
	SWL-100_{7day}	X			X						X					
	SWL-R_{7day}															
	SWL-F_{7day}															
	DWT_{mean}						O			O			O			
	DWT-30_{7day}			O			O		O	O			O			
Site C	PET		X	X		X	X		X	X		X	X		X	
	AP_{1day}															
	AP_{7day}	X			X				X							
	AP_{30day}		X			X					X					
	Q_{mean}							X					O			
	SWL_{mean}							X					O			
	SWL-25_{7day}					O						O			O	
	SWL-50_{7day}					O						O			O	
	SWL-75_{7day}					O						O			O	
	SWL-100_{7day}		X	O		O	O			O		O				X
	SWL-R_{7day}		X	X		X	X			X		X	X			O
	SWL-F_{7day}											O				X
	DWT_{mean}							X								
	DWT-30_{7day}															

Regarding the link between state variables and indicators of soil condition, at Sites A and C, SWE was correlated with SM and ST at the 95% significance level, while FD was only correlated with ST, in both cases over short cross-correlation distances (Figure 2.10). Conversely, at Site B, SWE and FD were found to have a statistically significant relation ($p < 0.05$) with SEC and SM, but only for the first three surveys following the onset of snowmelt.

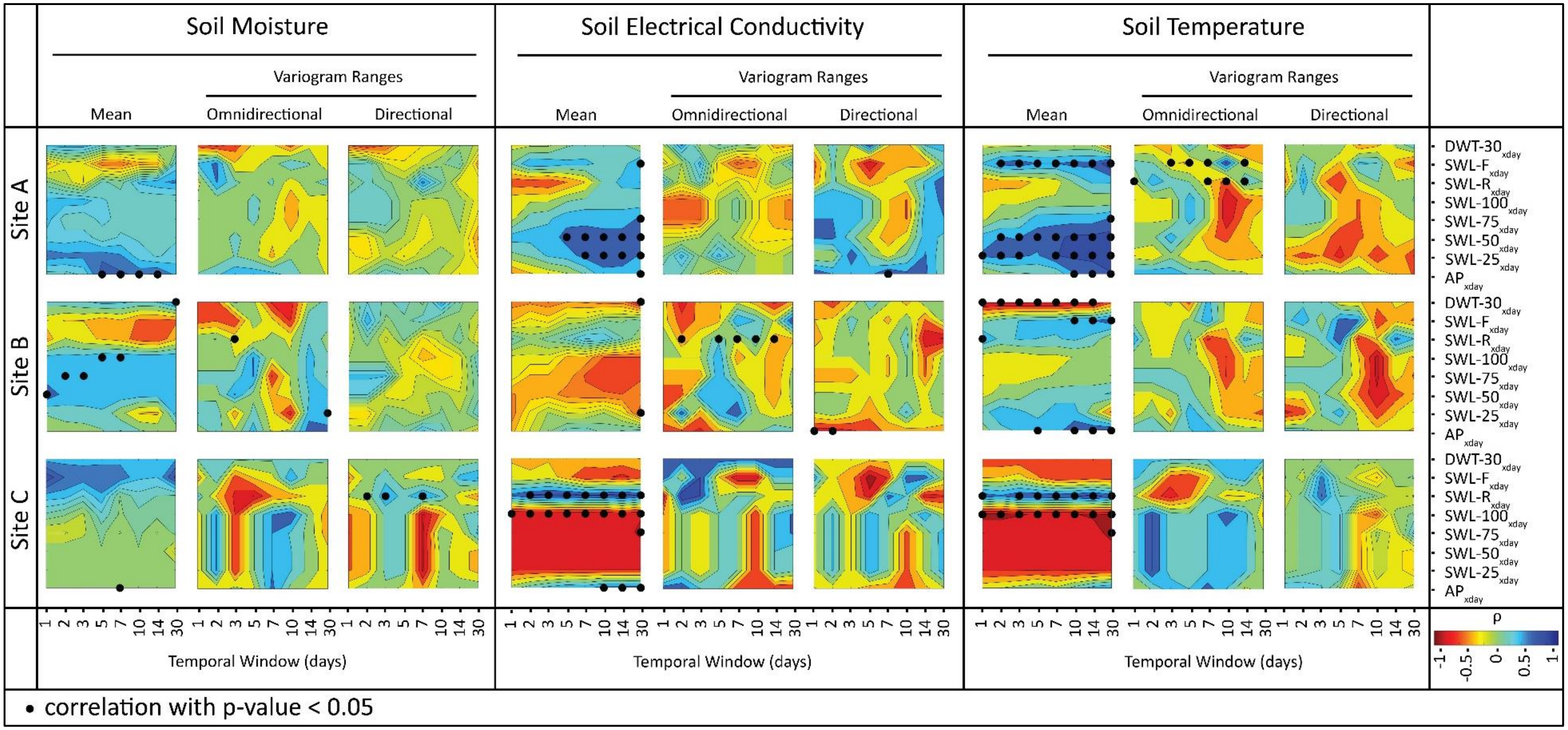

Figure 2.9. Summary of Spearman rank correlation analysis conducted between soil moisture, soil electrical conductivity and soil temperature pattern characteristics, surrogate measures of antecedent moisture conditions and indicators of streamflow response for multiple antecedent temporal windows. Descriptions of abbreviations can be found in Table 2.2 and in the text.

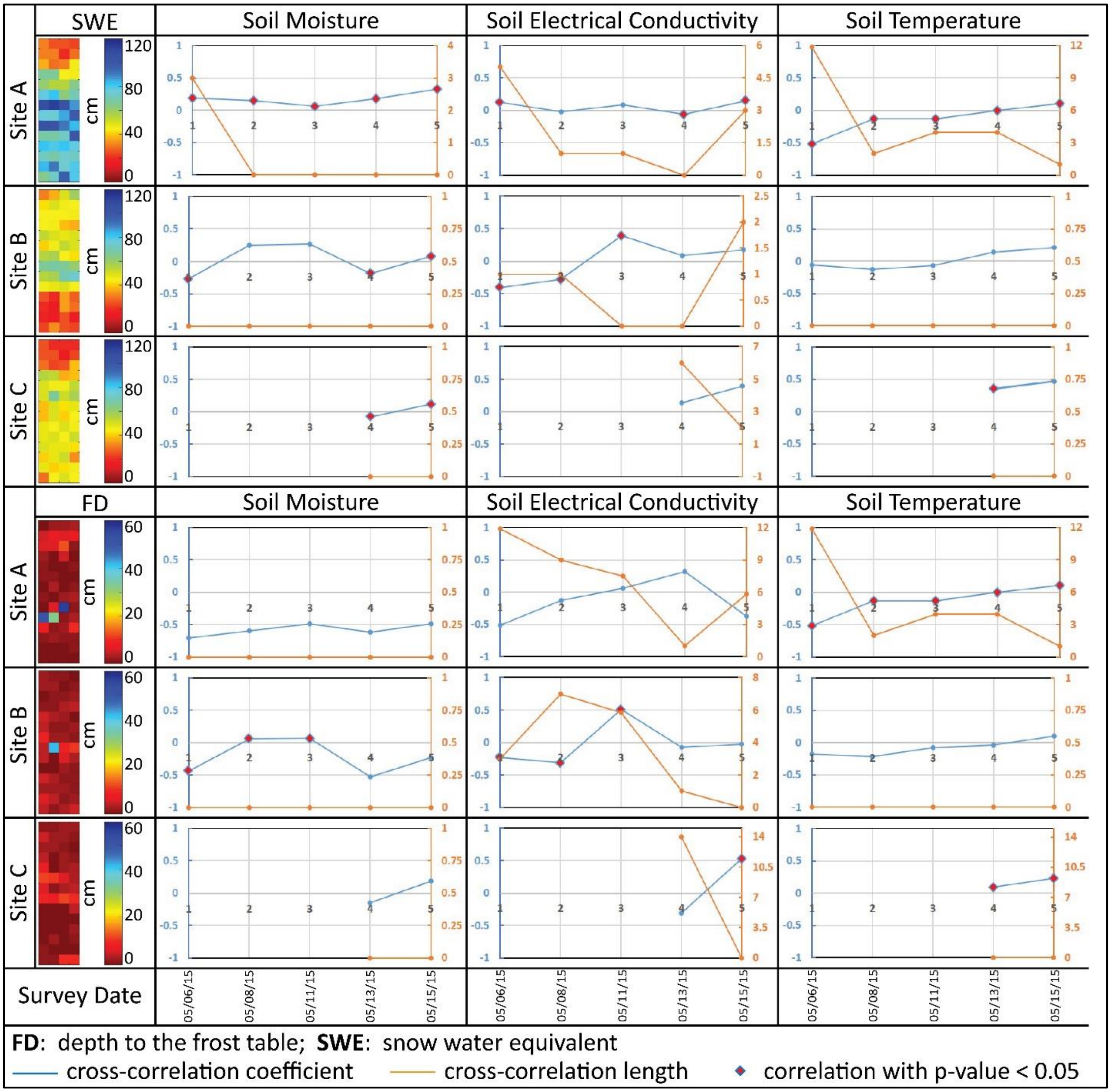

Figure 2.10. Similarity of state variable patterns (i.e., soil moisture, electrical conductivity and temperature) and soil condition patterns (i.e., snow water equivalent, depth to the frost table). Similarity is assessed through the cross-correlation coefficient and the cross-correlation length between two patterns. Here, SWE and FD are cross-correlated against soil moisture, soil electrical conductivity and soil temperature for the first 5 surveys following the 2015 spring snowmelt.

2.3.3 - Suitability of state variable pattern characteristics for predicting streamflow response

For SM, at all of the riparian area sites, the strongest statistically significant correlations involving Q_{mean} and SWL_{mean} were with site-specific and survey-specific minimum SM (Site A, max $\rho = 0.67$; Site B, max $\rho = 0.64$; and Site C, max $\rho = 0.73$) (Table 2.3 and Figure 2.9). However, non-spatial pattern statistics for SEC and ST were not correlated with Q_{mean} and SWL_{mean} at the 95% significance level. Indicators of streamflow response based on channel fullness values were correlated with non-spatial SM statistics at Sites A and B; the strongest correlations were between site-specific and survey-specific minimum SM and SWL-25_{5day} (ρ = 0.80) and between mean and median SM and SWL-50_{1day} (ρ = 0.75). Non-spatial SEC statistics and indicators of streamflow response based on channel fullness values were only correlated at Sites A and C; the strongest correlations were between minimum SEC conditions and SWL-25_{10day} (ρ = 0.84) and between mean SEC conditions and SWL-R_{3day} (ρ = 0.92). Those same indicators were correlated with non-spatial ST statistics at all sites; the strongest correlations at Site A were between mean ST and SWL-25_{30day} (ρ = 0.94); at Site B between maximum ST and SWL-F_{30day} (ρ = 0.78); and at Site C between mean ST and SWL-75_{30day} (ρ = -0.93). In contrast, statistically significant relations between spatial statistics and indicators of streamflow response were not found for any state variable or site.

2.4 Discussion

2.4.1 - Evidence of 'homogeneous' and 'heterogeneous' patterns in Prairie riparian areas

The spatial organization of SM has been used extensively to infer and distinguish differences in SM conditions (Ali and Roy, 2010a; Grayson et al., 1997; Teuling and Troch, 2005). In the CCW, wet and homogeneous SM conditions were perennially observed at Site B (Figure 2.5), while SM conditions at Sites A and C (Figures 2.4 and 2.6) were drier and heterogeneous unless higher-than-normal surface water levels were observed in the adjacent drainage channels, in which case wet and homogenous patterns were seen. Switching from dry to wet states was observed May 13-22 at Sites A and C (Figures 2.4 and 2.6): max AP_7 during this period was 42.0 mm at Site A and 49.0 mm at Site C, while the average AP_7 values for all other surveys at those sites were 9.1 mm and 8.1 mm.

SEC is generally positively correlated with SM, since dry soil is a poor conductor (Brady et al., 1996; Kachanoski et al., 1988). Accordingly, at Sites A and C, SEC patterns became more homogeneous with wetter conditions (Figures 2.4 and 2.6); however, unlike SM, SEC patterns never switched from heterogeneous to homogeneous. In soils within or adjacent to agricultural areas, the temporal variability of SEC has been largely attributed to water loss via evapotranspiration (Corwin and Lesch, 2005) and in the current study, a strong correlation between non-spatial SEC statistics and PET was found at all sites (Table 2.3). It has been shown that the spatiotemporal heterogeneity of SEC is controlled by static factors (e.g., clay content and mineralogy) and/or dynamic factors (e.g., soil water content, salinity and temperature) (Corwin and Lesch, 2005), with temporally persistent SEC patterns usually associated with static factors (Corwin and Lesch, 2005). In the CCW, we observed low temporal persistence in SEC at the point-scale, but high temporal persistence in the overall patterns. Correlations between SEC statistics and PET suggest that dynamic factors control point-scale SEC fluctuations while static factors are causing relatively time-invariant riparian SEC patterns.

ST is usually highest on bare soils but its variability also depends on aspect, vegetation, slope and water content, colour and texture of soil (Kang et al., 2000). In the CCW, ST at Sites A and C (Figures 2.4 and 2.6) – where there is not a developed canopy – increased more rapidly throughout the study period and reached higher temperatures in summer relative to Site B (Figure 2.5). Riparian ST patterns could not be categorized into distinct states as was the case with SM. However, in general, ST patterns at the homogeneously and continuously wet Site B were relatively constant and more spatially organized than was seen at the other sites.

2.4.2 - Influence of AMCs on state variable characteristics

The influence of AMCs on state variable characteristics and streamflow response has been shown to be highly variable in space and time (Grayson *et al.*, 1997; D’Odorico and Porporato, 2004;

James and Roulet, 2007; Ali and Roy, 2010b). Specifically, correlations between SM and indicators of AMCs vary depending on the chosen antecedent period (Ali and Roy, 2010b). In this study, several AMC measures based on different meteorological data and antecedent windows were correlated with non-spatial state variable statistics at all sites. For instance, a statistically significant relation between state variables and evaporation was expected – and found (Table 2.3) – as a result of surface and atmospheric interactions (Philip, 1957).

At Sites A and C, SM statistics were correlated with medium-term AMCs while ST statistics were correlated with mid- to long-term AMCs. SEC statistics were rather correlated with long-term AMCs and short-term AMCs at Site A and Site C, respectively. While there was a lack of consistency in the antecedent windows over which meteorological controls on SEC and ST were the strongest, the fact that medium-term AMCs were the most influential on SM indicates that soil wetness was not persistent in the long-term. Decreases in DWT at Sites A and C corresponded with intermittently wet periods (Figures 2.4 and 2.6). However, state variable statistics were mostly correlated with antecedent precipitation (AP_x) variables at Sites A and C, and to DWT variables at Site B. DWT_{mean} and AP_x were not always inter-correlated. In-fact, DWT_{mean} was only moderately correlated with AP_1 and AP_2 at Site B, and AP_5 and AP_7 at Sites A and C. Stronger correlations involving AP_x measures at Sites A and C are possibly a result of limited vegetative cover, partially or completely engineered drainage and better drained soils, relative to Site B.

At Sites A and C, spatial SWE patterns were most similar to SM patterns (Figure 2.10) but no influence of FD on state variables could be detected. This might be attributed to our methodology, as FD conditions changed very quickly during the early spring season and only a simplistic measure of FD was used in this study. Not considering frost type or measuring SM at shallower depths than 30 cm in early spring may have contributed to our inability to detect any interaction between FD, SM, SEC and ST.

2.4.3 - Suitability of riparian state variable pattern characteristics for predicting streamflow response

Previous studies have identified a link between SM and watershed response (Grayson et al., 1997; Western et al., 2004). In the CCW, minimum riparian SM had the strongest relation with SWL and Q, especially at Sites A and C (Table 2.3). Indicators of streamflow response dependent on channel fullness were correlated with state variable characteristics at all sites (Table 2.3 and Figure 2.9). Riparian SEC and ST have not traditionally been used as predictors of streamflow response but riparian ST has been associated with fluctuations in dissolved organic carbon concentrations that are strongly correlated with Q (Winterdahl *et al.*, 2011). SEC has also been used to evaluate influences on runoff (Friedman, 2005) and to identify runoff processes (Hümann *et al.*, 2011). Here we found SWL and Q to be correlated with SEC and ST less consistently than with SM, suggesting that riparian SEC and ST patterns are not good predictors of adjacent streamflow response. Very few statistically significant correlations were found between indicators of streamflow response and spatial statistics (variogram ranges for SM, SEC and ST patterns) (Figure 2.9). Other studies have shown inconsistencies in relations between watershed response and non-spatial versus spatial SM pattern statistics (Zehe *et al.*, 2005). In our case, non-spatial statistics were better predictors of streamflow than spatial statistics, likely because spatial organization was generally time-invariant: Site B was consistently wet while Sites A and C were “stuck” in a dry state unless a high precipitation threshold was exceeded, allowing wet patterns to develop.

2.4.4 - Process inferences

The identified relations between hydrological state variables, AMCs and streamflow response provide new insights into soil water movement in eastern Prairie riparian areas. We suspect that at Sites A and C, precipitation is generally able to infiltrate and move vertically through the soil profile. These sites have relatively well drained soils with high infiltration capacity (Fang et al., 2007), hence the typically dry SM patterns (Figures 2.4 and 2.6) and the strong correlations between non-spatial SM statistics and medium-term AP_x values (Figure 2.9). At Site B, the persistently homogeneous SM patterns and shallow DWT point toward lateral shallow subsurface flow and potentially saturation-excess overland flow. At Sites A and C, dry and wet SM conditions are likely associated with vertical and horizontal water movement respectively, thus confirming the preferred states hypothesis (Grayson *et al.*, 1997) at those sites.

Unlike Sites A and C, uniformly wet conditions combined with relatively constant and shallow DWT at Site B (Figure 2.5) suggest wetland-like dynamics that do not conform to the preferred states hypothesis. Some wetlands never flood (i.e., they never have a surface open water component) but have saturated upper soil horizons (hydric soils; see Brady *et al.*, 1996; Vepraskas *et al.*, 1997). Both hydric Aquolls soils and a high water table were observed at Site B (Figure 2.5), in addition to evidence that Site B may be a groundwater discharge area. Indeed, ST was higher in early spring and lower in the summer, a seasonal variation only observed at Site B and likely reflecting groundwater movement and discharge at that site (Cartwright, 1974). Furthermore, given that fluctuations in groundwater temperature are usually dampened by seasonal change, it is likely that the stable and homogenous ST conditions at Site B are a result of groundwater discharge. SEC maps from Site B also suggest the presence of groundwater ridging as elevated SEC was observed in the central portion of the site (Figure 2.5), potentially as a result of soil being in prolonged contact with a relatively saline groundwater source. The fact that SM, SEC and ST statistics at Site B were more strongly correlated to DWT measures then AP_x variables also hints at long memory effects at Site B, with state variable patterns being strongly influenced by “dead” storage and the long-term history of inputs (recharge) and outputs (discharge) and not so much by recent rainfall events. The main goal of the current study was to use hydrologic state variables to infer vertical and lateral subsurface water movements in eastern Prairie riparian areas. While some have indicated that SM is an indicator of subsurface flow (Grayson et al., 1997; Meyles et al., 2003; Western et al., 2004), others have rather argued that even when SM co-varies with streamflow, it is transient saturation at the soil-bedrock interface or near a soil layer of reduced permeability that is the real control (van Meerveld and McDonnell, 2005). In our study, organized soil moisture patterns at all sites could be linked to shallow subsurface lateral flow, especially in wet periods. SM therefore appeared to be a good indicator of shallow lateral subsurface water movement but less so as a predictor of adjacent streamflow response.

In general, the soil types present at each site agreed with the proposed interpretations. At Site A and C, dry conditions dominated with a temporary shift to the wet state following exceptionally large rainfall events. Well drained Mollisols at Site A can be associated with a variety of moisture regimes, thus our findings are not atypical (Soil Science Society of America, 2016). At Site C, wet Histols or Hemists, developed on glacial till or modified using engineered drainage, exhibited greater DWT and drainage potential (Soil Science Society of America, 2016). At Site B, the observed wetland-like dynamics are typical of Aquolls soils characterized by an aquic moisture regime resulting from saturation by ground water or by water of the capillary fringe (Vepraskas *et al.*, 1997).

2.5 Conclusion

In this chapter we characterized non-spatial and spatial characteristics of riparian SM, SEC and ST patterns as well as the relations between those state variables, indicators of AMCs and streamflow response in a near-level Prairie watershed. SM patterns at two sites were shown to switch between two preferred states, a wet and a dry state. However, such distinct states could not be identified for SEC and ST. SM, SEC and ST patterns combined with DWT data were used to infer soil water movement such as groundwater ridging at a headwater, forested riparian area that exhibited wetland-like hydrologic dynamics. Analyses also confirmed that PET, DWT and AP_x have a significant influence on the characteristics of riparian state variables; however, the usefulness of all three state variables for predicting streamflow response in adjacent channels was not fully confirmed.

The outlined analyses and inferences suggest that in addition to soil texture, vegetation and water drainage characteristics, spatial and non-spatial characteristics of riparian SM, SEC and ST can be useful for inferring riparian-to-stream water movement in near-level landscapes. Further investigations are however necessary to confirm the usefulness of SM spatial patterns for predicting streamflow response in other landscapes across the Canadian Prairies.

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CHAPTER 3.

PRELIMINARY SYNTHESIS AND TRANSITION

The analyses reported in the previous chapter led to the identification of a soil moisture 'switching behavior' for two of the monitored riparian areas between two preferred states, namely a wet and a dry state. However, this behaviour was not observed for the other hydrologic state variables that were considered, namely soil temperature and soil electrical conductivity. In some circumstances, patterns of soil moisture, soil electrical conductivity and soil temperature in combination with water table data could be used to infer subsurface water movement. The last chapter also allowed the confirmation of significant influence of potential evaporation, water table levels and antecedent precipitation on riparian hydrologic state variables.

The literature review included in Chapter 1 highlighted the fact that the high spatial variability in physical characteristics and hydrologic response across the Prairies makes it challenging to identify dominant runoff generation mechanisms, which are known to be influenced by a variety of factors. The analyses carried in Chapter 2 confirmed that even while focusing on three sites only, consistent hydrological responses and control factors on those responses could not be identified. The difficulty in precisely identifying dominant runoff processes could be due to the inherent variability of said processes or to the fact that indirect data was used. Indeed, while spatial patterns of SM, SEC and ST are useful to assess the variability in soil water conditions across the three monitored riparian areas, movements of soil water were not measured directly and could only be inferred from the state variable patterns. The significant temporal changes in the state variable patterns however hint at potentially significant soil water movement, thus raising the question of whether shallow subsurface flow is a significant runoff generation mechanism in near-level Prairie landscapes. The remainder of this thesis therefore builds upon Chapter 2 via the use of tracer data to track, in a more direct manner, the movement of surface and subsurface water from riparian areas to drainage channels. Chapter 4 focuses on evaluating streamflow composition during different types of hydrological events (i.e., snowmelt- or rainfall-triggered) and linking streamflow response to snow cover dynamics, antecedent moisture conditions, and flow velocity characteristics of both subsurface and surface water.

CHAPTER 4.

EVALUATING THE RELATIVE IMPORTANCE OF SHALLOW SUBSURFACE FLOW IN A NEAR-LEVEL PRAIRIE LANDSCAPE

4.0 Abstract

The hydrologic literature on the Northern Great Plains, including the Canadian Prairies, has mainly focused on Hortonian overland flow as the dominant runoff generation mechanism. The main objective of the current study was therefore to shift the focus to shallow subsurface flow and assess whether there are instances when it generates the majority of runoff in relatively flat Prairie landscapes. Three "riparian-to-stream" sites with contrasted land cover were monitored within the Catfish Creek Watershed (south-eastern Manitoba, Canada), and four hydrologic events (snowmelt-triggered and rainfall-triggered) were selected for detailed analysis. A dual water sourcing strategy, relying on conservative ($\delta^{18}O$) and non-conservative (fluorescent dyes) tracers, hydrograph separation and breakthrough curves, was applied in order to distinguish the relative importance of surface ('new' water) and subsurface ('old' water) runoff sources. Surrogate measures of antecedent moisture conditions (AMCs) as well as data on streamflow and water table fluctuations were also used to identify the main factors influencing the contributions of different runoff sources to streamflow. Results showed significant 'old' water contributions during both snowmelt- and rainfall-triggered events, albeit subsurface runoff sources appeared to be especially significant during the summer season. Decreases in snow cover extent were accompanied by 'flashy', high magnitude stream hydrographs at an engineered grassland site, but not at a naturally vegetated grassland site or a headwater forested site. Larger peak discharge was observed coincident to elevated AMCs at all sites and in general, surface flow velocities were faster than subsurface flow velocities, with the exception of one rainfall-triggered event. A rapid delivery of 'old' water to streams was generally observed and hypothesized to be the result of macropore flow, piston flow or transmissivity feedback depending on the season and the position of the water table. Further investigations would be useful to better understand the spectrum of subsurface runoff generation mechanisms that may promote the rapid movement of 'old' water from riparian areas to streams in relatively flat landscapes.

Keywords

Canadian Prairies, hydrograph separation, antecedent moisture conditions, stable water isotopes, fluorescent dyes, breakthrough curves, subsurface flow, flow velocities

4.1 Introduction

Understanding how precipitation is transformed into runoff is critical in developing hydrologic conceptualizations and models (Davie, 2008), and an extensive history of research focusing on the relative importance of specific flow pathways in hillslope environments exists (e.g., Hewlett and Hibbert, 1967; Dunne and Black, 1970; Kirkby, 1988; Bonell, 1993). Past research efforts revealed the existence of several runoff generation mechanisms that activate when a specific combination of characteristics related to the precipitation event, antecedent soil moisture (SM) conditions, other soil characteristics, topography and land use are present (Anderson and Burt, 1978; Bonell, 1993; Bracken and Croke, 2007). In particular, since the 1970s, subsurface flow has consistently been identified as a runoff generation mechanism with greater contributions to streamflow than previously expected (e.g., Anderson & Burt, 1978; Dunne & Black, 1970; Sklash & Farvolden, 1979; van Meerveld & McDonnell, 2005; Weiler, 2005). The fact that long-term research has been conducted in iconic hillslope environments (e.g., Coweeta, North Carolina; Sleepers River, Vermont; Hubbard Brook, New Hampshire; and Panola Mountain, Georgia) and that water is typically delivered rapidly to streams from steep areas (Beven, 2006) is partially responsible for the numerous hillslope-focused conceptualizations of runoff generation. Few similar studies have been conducted in high-latitude, sub-humid to arid environments and relatively flat regions like the Canadian Prairies (hereafter referred to as the Prairies), thus causing a need for runoff conceptualizations suitable to seasonally dry environments without typical, steep hillslopes.

Runoff generation mechanisms (or flow pathways) are traditionally classified into surface and subsurface mechanisms. For surface pathways, Hortonian overland flow (HOF) dominates when water input intensity exceeds infiltration capacity (Horton, 1933; Black, 2005), while saturation-excess overland flow (SOF) prevails when precipitation falls onto (or subsurface water exfiltrates from) previously saturated soils (Hewlett and Hibbert, 1967). For subsurface pathways, regolith subsurface flow and groundwater flow refer to water movement through the vadose zone and phreatic zone, respectively (Dingman, 2015). Direct measurement of these flow pathways is however not straightforward as precipitation and streamflow data are relatively easy to obtain but overland and subsurface flow are more elusive to quantify, especially since variability in physical and climatic factors makes it challenging to obtain representative measurements at large scales. Besides, those direct measurements are often insufficient to identify both temporal and geographic (or spatial) sources of water: the former are distinguished based on the age of the water while the latter rely on the physical origin of the water, either laterally from a location along the hillslope-riparian area-stream continuum or vertically from a specific depth in the vadose or phreatic zone (McGlynn and McDonnell, 2003; Sayama and McDonnell, 2009). Alternatively, investigation of surface and subsurface flow sources at the watershed scale can be pursued using the geochemical signature of stream, precipitation and subsurface water (Sklash et al., 1976; Sklash and Farvolden, 1979; Christophersen et al., 1990; Hooper et al., 1990; Robson et al., 1992; Bonell, 1993; Buttle, 1994; Genereux and Hooper, 1998). Indeed, each flowpath potentially exposes water to different geologic material, resulting in distinct travel times, velocities and geochemical and isotopic signatures that provide a means to study hydrological processes at a range of scales (Sklash, 1990). Those signatures – which are natural tracers – have been used to examine shallow subsurface flow and partition stream water into temporal sources, namely pre-event ('old') and event ('new') water through hydrograph separation (Hooper and Shoemaker, 1986; Wels et al., 1991; Klaus and McDonnell, 2013). Artificial tracers such as fluorescent dyes have also been applied to small areas to evaluate preferential flow pathways (e.g., cracks and macropores; Beven and Germann, 1982; Guebert and Gardner, 2001; Weiler and Naef, 2003; Weiler and Flühler, 2004) and to trace the movement and rate of movement of deep ground water (Smart and Laidlaw, 1977), shallow subsurface water (Pang et al., 1998), surface runoff (Abrahams et al., 1986), snowmelt (Campbell et al., 2006) and infiltrated water into thawed and frozen soils (Stadler et al., 2000).

The combination of hydrometric and tracer data to aid hydrologic conceptualization has been recently promoted (e.g., Klaus and McDonnell, 2013) and would be of particular help in low-relief and seasonally dry environments where runoff events are transient and difficult to capture. Prairie landscapes are one example of such environments and cover a significant portion of the North American landmass, with 520,000 km^2 in Canada and 780,000 km^2 in the United States. The western portion of the Prairies is undeniably the most well-known with its numerous small lakes and sloughs (i.e., pothole wetlands), and regional hydrologic conceptualizations and models for this region have traditionally focused on HOF and surface fill-and-spill processes (e.g., Fang et al., 2007; Granger et al., 1984; Shaw et al., 2013; Zhao & Gray, 1999). Eastern Prairie landscapes, however, are vastly different from the Prairie Pothole Region in the west and can be seen as good archetypical landscapes to investigate dominant runoff pathways in large-scale, low-relief, non-pothole-dominated, seasonally cold and seasonally dry watersheds. For such low-relief environments located at high latitudes, the primary hydrological event is often the spring snowmelt, typically characterized by HOF, i.e., rapid surface runoff travelling over frozen ground (Fang et al., 2007) due to the reduced soil infiltration capacity (Brady et al., 1996). Factors that influence meltwater infiltration, such as SM content, soil temperature, saturated hydraulic conductivity, soil texture and vegetative cover are highly variable throughout the Prairies (Zhao and Gray, 1999) and the snowmelt rate is often affected by extensive snow re-distribution and intermittent midwinter thaws (Granger et al., 1984; Fang et al., 2007, 2010). Measuring the rate of water input to streams during snowmelt is particularly challenging due to the uneven distribution of snow density and snow water equivalent (SWE) within a snowpack (van der Kamp et al., 2003; Fang and Pomeroy, 2009; Fang et al., 2010). As for summer rainfall events in the Prairies, they are often high-intensity convective storms occurring over a single day (Dyck and Gray, 1979; Shook and Pomeroy, 2012) which produce surface runoff only exceptionally (Shook and Pomeroy, 2012), due to the depletion of SM from evapotranspiration that leaves soils unsaturated for most of the summer (Brady et al., 1996; Fang et al., 2007). The significant presence of arable land in the eastern Prairies does make HOF possible, since infiltration-excess is known to occur on un-vegetated land with low infiltration rates (Mosley, 1979). A number of characteristics related to frost formation, soil depth and permeability and macropores however make it possible for shallow subsurface runoff contributions to be significant in the eastern Prairies, although the extent to which they might remain unclear. For instance, in wooded areas, granular and honeycomb frost has been shown to occur near the ground surface while concrete frost may appear at greater depths (Striffler et al., 1959). Generally, frozen ground results in reduced surface permeability (Black, 2005), especially where concrete frost has developed (Striffler et al., 1959). However, in some cases, honeycomb and granular frost may enhance infiltration (Trimble et al., 1958). Granular and honeycomb frost overlying concrete frost offer a potential mechanism for shallow subsurface runoff generation: melt water may infiltrate the soil, eventually reaching a concrete frost impeding layer, thus triggering lateral preferential flow along the soil-impeding layer interface. Further potential for subsurface runoff contributions is allowed by desiccation cracks and macropores during dry periods that may quicken the routing of meltwater and rainwater to the vadose zone. Depending on macropore connectivity and crack depth, meltwater can reach deeper soil layers faster, potentially even soil layers located below the frost table (Haupt, 1967; Harris, 1972). The south-eastern Prairies also have soils that are generally richer in organic deposits (peat) than soils in the western Prairies (Province of Manitoba, 2015), which impacts infiltration capacity, increases SM retention (Brady et al., 1996), and influences the development, type and persistence of frost (Trimble et al., 1958; Haupt, 1967; Harris, 1972; Granger et al., 1984; Fang et al., 2010). Besides, important human activities have the potential to enhance the occurrence and magnitude of both surface and subsurface runoff, either through agricultural activities – including soil compaction and tillage – that take place very close to streams in summer and fall, or because of artificially high water levels maintained in engineered drainage channels and that lead to persistent riparian flow reversals in some areas. The main goal of the current study was therefore to use both hydrometric and tracer data toward assessing whether shallow subsurface flow is a significant runoff generation mechanism in low-relief (or near-level) eastern Prairie landscapes that have not yet been the focus of extensive hydrologic conceptualization studies. Three specific research questions were considered, namely:

1) What are the main temporal sources (i.e., old versus new water) of Prairie streamflow and does their relative importance change depending on event type (i.e., snowmelt or rainfall)?
2) Beyond the effect of event type, do snow cover extent and antecedent moisture conditions influence event response characteristics as well as temporal and geographic sources of streamflow?
3) Are subsurface and surface water travel times to the stream significantly different from one another?

4.2 Methods

4.2.1 - Study site description

The mesoscale Catfish Creek Watershed (CCW) is located approximately 100 km north-east of the city of Winnipeg (south-eastern Manitoba, Canada, see Figure 4.1). It drains a total area of 642 km^2 into Lake Winnipeg and maintains a near-level topographic profile (maximum relief of 91.4 m, average slope of 1.4 degrees or 2.5 %), with land use equally split between agricultural and forested land. Natural watershed conditions have been altered extensively through wetland (peatland) drainage for agricultural use and the addition of road and drainage networks (Elliott, 1977). Geologically, the CCW is characterized by hummocky, crystalline Archean bedrock mantled by a discontinuous veneer of sandy glacial till with morainal uplands and extensive fen peatland (Smith et al., 1998). Natural waterways in the CCW are relatively young and are incised in deposits left by glacial activity (Teller and Leverington, 2004). The watershed typically experiences short, warm to hot summers and long, cold winters with an average annual temperature of 1.9°C and mean annual precipitation of roughly 530 mm with significant spatial and year-to-year variability (Environment Canada, 2011). One fifth of annual precipitation is delivered as snow and the watershed is characterized by an average annual moisture deficit of roughly 90 mm.

For the current study, three sites were selected in the CCW based on their contrasting characteristics that cumulatively represent the area (Figure 4.1). Given the low-relief topography associated with the presence of large and flat riparian areas and floodplains and the absence of typical hillslopes, the focus was on "riparian-to-stream" study sites to investigate the relative importance of surface and subsurface runoff generation. Site A is located adjacent to a man-made drainage channel, has an average slope ranging from 5 to 9 %, is naturally vegetated with a mixture of grass species and has well-drained Solonetzic Gray Luvisol soils (GIS4AG, 2015; Soil Science Society of America, 2016). Site B is a headwater forested site situated adjacent to a small headwater creek and is the most eastern site, located near the interface between the Boreal Prairies and Boreal Shield ecozones (Smith et al., 1998). This site has an average slope between 5-9% and has poorly drained Rego Humic Gleysol soils (GIS4AG, 2015). Site C is located in the southern portion of the CCW in a lowland primarily used for agriculture. This site is adjacent to an engineered drainage dyke and a man-made channel: it has an average slope of 0-2% and poorly drained Terric Messisol soils developed on mesic fen peat overlying loamy to clayey lacustrine sediments (GIS4AG, 2015).

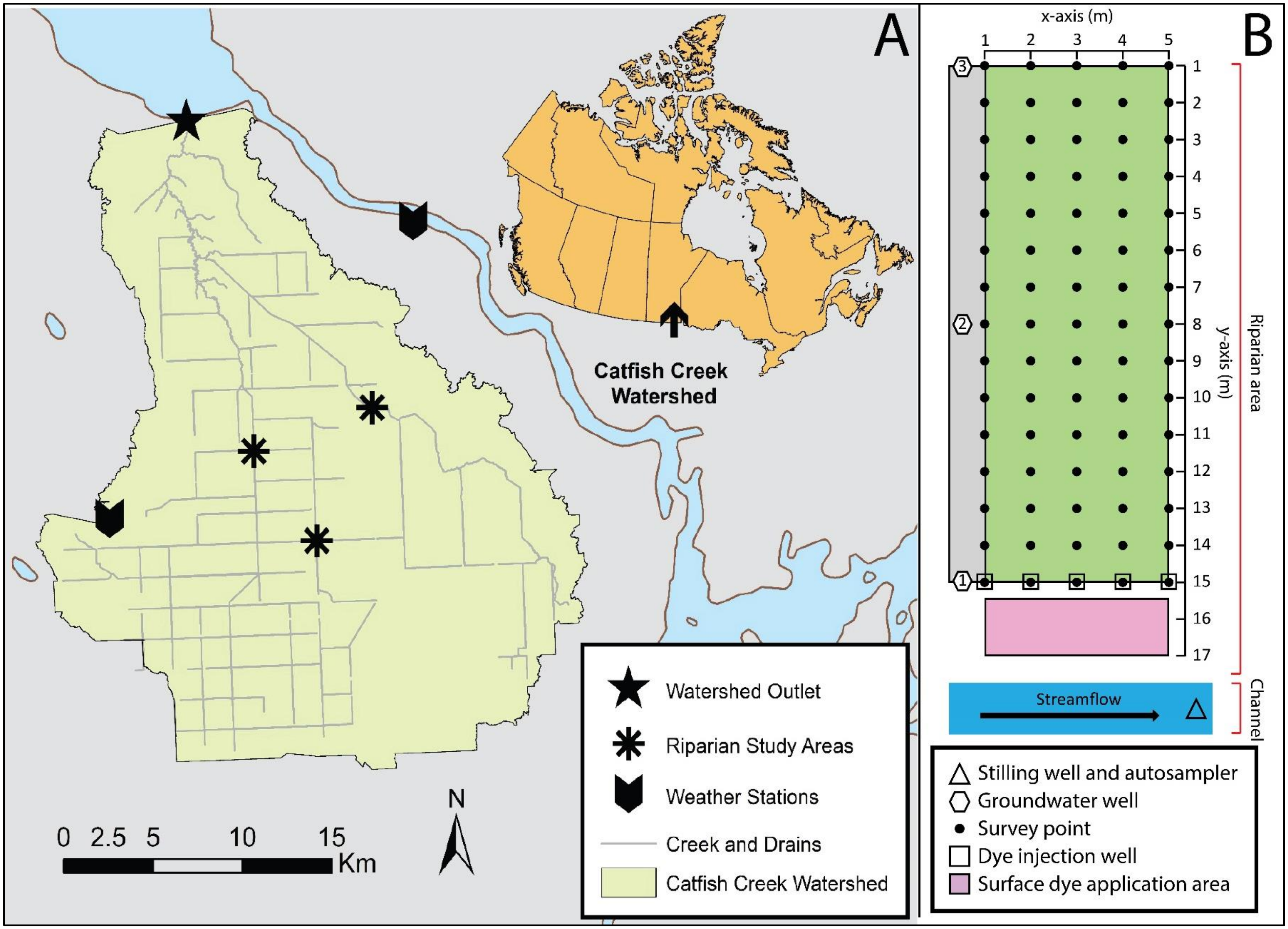

Figure 4.1. (A) Location of the Catfish Creek Watershed within Canada as well as the riparian areas and weather stations used in this study. (B) Experimental setup used at each riparian area. All groundwater and injection wells are 1.5 m deep. Groundwater wells and stilling wells were outfitted with capacitance-based water level loggers.

4.2.2 - Event-based hydrometric and tracer data collection

The three riparian study sites shared a common experimental setup consisting of a rectangular plot oriented perpendicularly to the adjacent channel (Figure 4.1). Plots featured a 5 m x 15 m grid with 75 equally spaced nodes for manual spatial surveys of key target variables (see below). Three groundwater wells (1.5 m deep) located on a plot edge perpendicular to the channel and a stilling well in the channel were installed at each site (Figure 4.1). Additionally, at each site, five injection wells (1.5 m deep) were deployed on a transect parallel (and proximal) to the adjacent drainage channel for dye-tracer experiments (see below).

Four hydrologic events (two snowmelt-triggered and two rainfall-triggered) were selected for monitoring in spring and summer 2015 (Table 4.1). Event monitoring included tracing experiments using fluorescent dyes: Rhodamine WT (RHWT) was applied over snow or bare soil to a designated 2-meter near-stream area using a pressurized handheld pesticide applicator (Figure 4.1), while water soluble Fluorescein (FL) was introduced in the five injection wells at each site (Figure 4.1), thus allowing the comparison of two geographic sources defined by soil depth (i.e., surface versus vadose zone). The dye concentrations and volumes used were relatively constant across sites and across events (or experiments), with the exception of an interruption during Event 2 for which vandalism close to Site C and technical issues at Site B resulted in missing data (Table 4.1). Overall, low dye concentrations were used to adhere to conditions established when permission to conduct the experiments was granted by Manitoba Conservation and Water Stewardship. Before and throughout each event and at each site, ground water, rain water and snowpack samples were collected to establish background dye tracer concentrations and document the isotopic signatures of ‘old’ and ‘new’ water sources. It is worth noting that here the term “ground water” was used to refer to both soil water in the vadose zone and water in the phreatic zone, given that the water table delimiting those zones is highly variable in time. A dual water sourcing strategy was used, with conservative ($\delta^{18}O$) and non-conservative tracers (fluorescent dyes), in an attempt to gather different information and seek result convergence. Indeed, dye tracers can be used to evaluate the course of water movement, hydrologic connectivity between riparian areas and adjacent stream channels, and to estimate water flow velocities at the event scale using dye breakthrough curves (Sabatini and Austin, 1991; Flury, 2003; Gerke et al., 2013). As for stable water isotopes such as $\delta^{18}O$, they are most commonly used to perform temporal source-based hydrograph separation (Sklash and Farvolden, 1979; Klaus and McDonnell, 2013). An electronically controlled, automated water sampler was located in the drainage channels adjacent to each experimental plot (Figure 4.1) and programmed to collect stream water samples – at frequencies ranging from two to twelve hours – before, throughout and after each event. All samples were tested for dye-tracer concentrations and stable water isotopic ratios ($\delta^{18}O$ and $\delta^{2}H$) using two Turner Design™ 10-AU fluorometers fitted with specific filters for RHWT and FL, and a liquid water isotope analyzer based on cavity ringdown spectroscopy (CRDS) technology. Minimum dye-tracer detection using the fluorometers was 0.01 ppb, and the instrument precision for determining stable water isotopic ratios was < 0.025 per mil and < 0.1 per mil for $\delta^{18}O$ and $\delta^{2}H$, respectively.

Table 4.1. Summary of event dates and fluorescent dye tracer experiment characteristics. Unforeseen circumstances (e.g., vandalism) resulting in missed experiments or missing components are indicated by 'NA'.

Event characteristics												
		Event 1			Event 2			Event 3			Event 4	
Event trigger		Snowmelt			Snowmelt			Rainfall			Rainfall	
Start date (mm/dd/yr)		03/12/15			04/01/15			05/14/15			07/23/15	
End date (mm/dd/yr)		03/21/15			04/12/15			05/31/15			07/28/15	
Event based dye tracer characteristics												
Site	A	B	C	A	B	C	A	B	C	A	B	C
Dye application (mm/dd/yr)		03/11/15			04/05/15			05/15/15			07/23/15	
Rhodamine volume (mL)	1600	1600	1600	885	NA	NA	2000	2000	2000	450	450	450
Fluorescein volume (mL)	1000	1000	1000	1000	1000	NA	1000	1000	1000	1000	1000	1000
Rhodamine [ ] (ppb)	150	150	150	150	NA	NA	150	150	150	150	150	150
Fluorescein [ ] (ppb)	140	140	140	140	140	NA	140	140	140	140	140	140
Event based sample count												
Site	A	B	C	A	B	C	A	B	C	A	B	C
# of samples for dye tracer concentration and stable water isotopic ratio analyses	14	14	12	41	37	NA	26	25	25	47	47	47

To complement tracer data, spatial surveys of snow depth (SD) were completed at each site before and during snowmelt-triggered events (Table 4.1) by using a measuring stick from the ground surface to the top of the snowpack at each of the 75 grid nodes. At the same time, snow water equivalent (SWE) was measured but only at the four corners of the survey grids: this was done to avoid snow coring within the grid and hence disturbing the snowpack in a major way. Site-specific average snow density was derived from those SWE measurements and used to convert the 75-point SD survey maps to SWE maps. Spatial surveys of soil moisture (SM) were completed before and during rainfall-triggered events (Table 4.1) using a capacitance-based probe (AQUATERR Instruments and Automation) at a depth of 30 cm, with values ranging from 0 (dry soil) to 100% (saturated soil). Capacitance-based water level loggers (Odyssey^{TM}) were deployed in the groundwater wells and stilling wells and used to record subsurface water levels (WTB, expressed as the height of water, in mm, above the base of the wells) and stream water levels (SWL) every 15 minutes. Site-specific rating curves were used to convert SWL values to discharge (Q) and develop flow hydrographs. For each timestep, the depth to the water table (DWTB) was also obtained by 1500 – WTB, where 1500 is the depth of the groundwater wells, in mm.

Event-specific surrogate measures for antecedent moisture conditions (AMCs) were computed based on rainfall, SM and SWE data. Rainfall data was obtained from two weather stations (Figure 4.1) and for each site, inverse distance weighting (Price et al., 2000) was applied to account for the distance between each site and the weather stations. Snowmelt was estimated by computing the difference in snow depth between consecutive surveys and assuming that losses via sublimation were negligible. Antecedent precipitation (AP) measures (i.e., cumulative rainfall and snowmelt for the days preceding each event) were computed over a range of antecedent temporal windows (e.g., x days before each event, with x varying between 1 and 30) in order to characterize short-, medium- and long-term meteorological influences.

4.2.3 - Data processing and analysis

Using isotopic tracer data ($\delta^{18}O$ only), hydrograph separation was performed for each stream water sample collected. For the sake of simplicity, it was assumed that all subsurface runoff (shallow flow over the first 1.5 m of the soil profile) was associated with the 'old' water signature while surface runoff was associated with the 'new" water signature. First, data yielded from isotopic analysis was plotted against both the global (GMWL: $\delta^{2}H = 8^{*}\ \delta^{18}O + 10$) and local (LMWL: $\delta^{2}H = 7.78^{*}\ \delta^{18}O + 6.22$) meteoric water lines to ensure that fractionation or sample degradation had not significantly impacted data quality. The basic mass balance equations underlying the hydrograph separation approach were then considered:

$$Q_t = Q_{old} + Q_{new} \qquad \text{(Eq. 1)}$$

$$Q_t\,\delta_t = Q_{old}\delta_{old} + Q_{new}\delta_{new} \qquad \text{(Eq. 2)}$$

where Q is discharge, δ refers to isotopic ratios of a sample relative to a standard, and subscripts t, old and new refer to total, 'old' water and 'new' water (Bazemore et al., 1994). After transformation of those equations, the proportions of old water ($PROP_{old}$) and new water ($PROP_{new}$) were calculated as:

$$PROP_{old} = \frac{\delta^{18}O_t - \delta^{18}O_{new}}{\delta^{18}O_{old} - \delta^{18}O_{new}} \qquad \text{(Eq. 3)}$$

$$PROP_{new} = 1 - PROP_{old} \qquad \text{(Eq. 4)}$$

where the new water signature was either estimated using a snow sample, a rain water sample or an average of the two when rain-on-snow occurred. The old water signature was rather estimated using groundwater samples (Burns et al., 2001).

Event dynamics were characterized using a range of summary statistics for precipitation, water table behavior, streamflow, old water proportions and dye breakthrough curves. Precipitation statistics included site-specific AP (see above), total event rainfall in addition to the available SWE, 7-day melted SWE and total event SWE. Total rainfall was simply a cumulative sum of rainfall over the defined event duration; total available SWE was the average snowpack thickness expressed as SWE preceding the event; 7-day melted SWE was the sum of melted snow for 7 days preceding the event; and total event SWE was the total amount of snow that melted over the event duration. Riparian water table and streamflow dynamics were characterized via event peaks in Q, channel fullness, subsurface water levels (minimum DWTB) and the maximum increase (rise) in WTB. Channel fullness was expressed as the percentage of the channel filled with water, and was determined by dividing water levels recorded at each stilling well by the site-specific drainage channel depth. Additionally, the time elapsed from the start of the event to the hydrograph or WTB initial rise (lag to rise), and from the start of the event to the hydrograph or WTB peak (lag to peak) was determined. In some cases, stream water levels (SWL) were used to visually represent streamflow rather than Q since site-specific rating curves (based on exponential relationships) sometimes resulted in very large Q magnitudes that masked smaller, short-term fluctuations that are visible in the SWL timeseries.

For each event, the variability in 'old' water contributions to streamflow was assessed using the mean, median, minimum, maximum, standard deviation, and coefficient of variation of $PROP_{old}$ values, the number of peaks in the $PROP_{old}$ timeseries across the duration of the event, as well as the lag to the initial $PROP_{old}$ rise and lag to the highest $PROP_{old}$ value. Concentrations of RHWT and FL in stream water samples were characterized with the same summary statistics as $PROP_{old}$ and the same timing parameters (i.e., lag to rise and lag to peak). Additionally, average and maximum travel velocities were estimated for both dye tracers: average travel velocities were computed as the distance of travel (between the application area to the stream sample collection point) divided by the lag to peak dye concentration, and maximum travel velocities as the distance of travel divided by the lag to initial rise in dye concentration.

In relation to research question 1 (i.e., *what are the main temporal sources (i.e., old versus new water) of Prairie streamflow and does their relative importance change depending on event type (i.e., snowmelt or rainfall)?*), $PROP_{old}$ characteristics (e.g., median, maximum, lag to rise and lag to peak) for snowmelt-triggered and rainfall-triggered events were compared using the Kruskal-Wallis test (McKight and Najab, 2010) to see if they were significantly different from one another. The Kruskal-Wallis test examines whether two or more independently sampled groups of data came from the same distribution; the alternative hypothesis is that not all samples come from the same distribution. p-values less than 0.05 mean that the null hypothesis can be rejected, which here would mean that snowmelt- and rainfall-triggered events are significantly different from one another.

Graphical representations were used to evaluate the influence of snow cover extent and AMCs on event response characteristics and temporal and geographic sources of streamflow (research question 2). Precipitation, pre-event and event SWE and SM data were paired with DWTB, streamflow and $PROP_{old}$ timeseries and dye-tracer breakthrough curves for qualitative interpretation.

Determining if there was a significant difference in travel times for water delivered to the stream via subsurface and surface pathways (research question 3) was done, in part, through the comparison of average and maximum travel times using the Kruskal-Wallis test. All statistical analyses were performed using MATLAB (R2015). Lastly, the movement of subsurface water at the event scale, especially its direction from the riparian area to the adjacent drainage channel or vice versa, was examined using summary statistics to describe the variability in riparian hydraulic gradients at each site. For each event timestep, the riparian hydraulic gradient was computed as:

$$\text{Hydraulic Gradient} = \frac{\text{WTB elevation at GW3} - \text{WTB elevation at GW1}}{\text{Distance between GW1 and GW3}} \qquad \text{(Eq. 5)}$$

where WTB elevation was in fact elevation-corrected WTB data expressed relative to sea level (Dingman, 2015).

4.3 Results

4.3.1 - Temporal sources of streamflow during snowmelt- and rainfall-triggered events

Riparian groundwater and SWL dynamics were highly variable temporally and between sites throughout the 2015 observation period (Figure 4.2). While high synchronicity between SWL and DWTB was typically observed at Sites A and C, short-term fluctuations in SWL were seldom reflected in DWTB timeseries at Site B. Occasionally, fully saturated soils resulted in surface water ponding at Sites A and B, but not at Site C. The ‘flashiness’ of event response was also site-dependent: peaks in SWL appeared to be sharper (i.e., narrower) at Sites B and C compared to Site A. A graphical comparison between the event-specific isotopic signatures of the collected water samples and both the GMWL and LMWL showed no significant deviations (Figure 4.3), indicating those samples had not suffered any significant fractionation-related deterioration and could be used for further analysis (i.e., hydrograph separation).

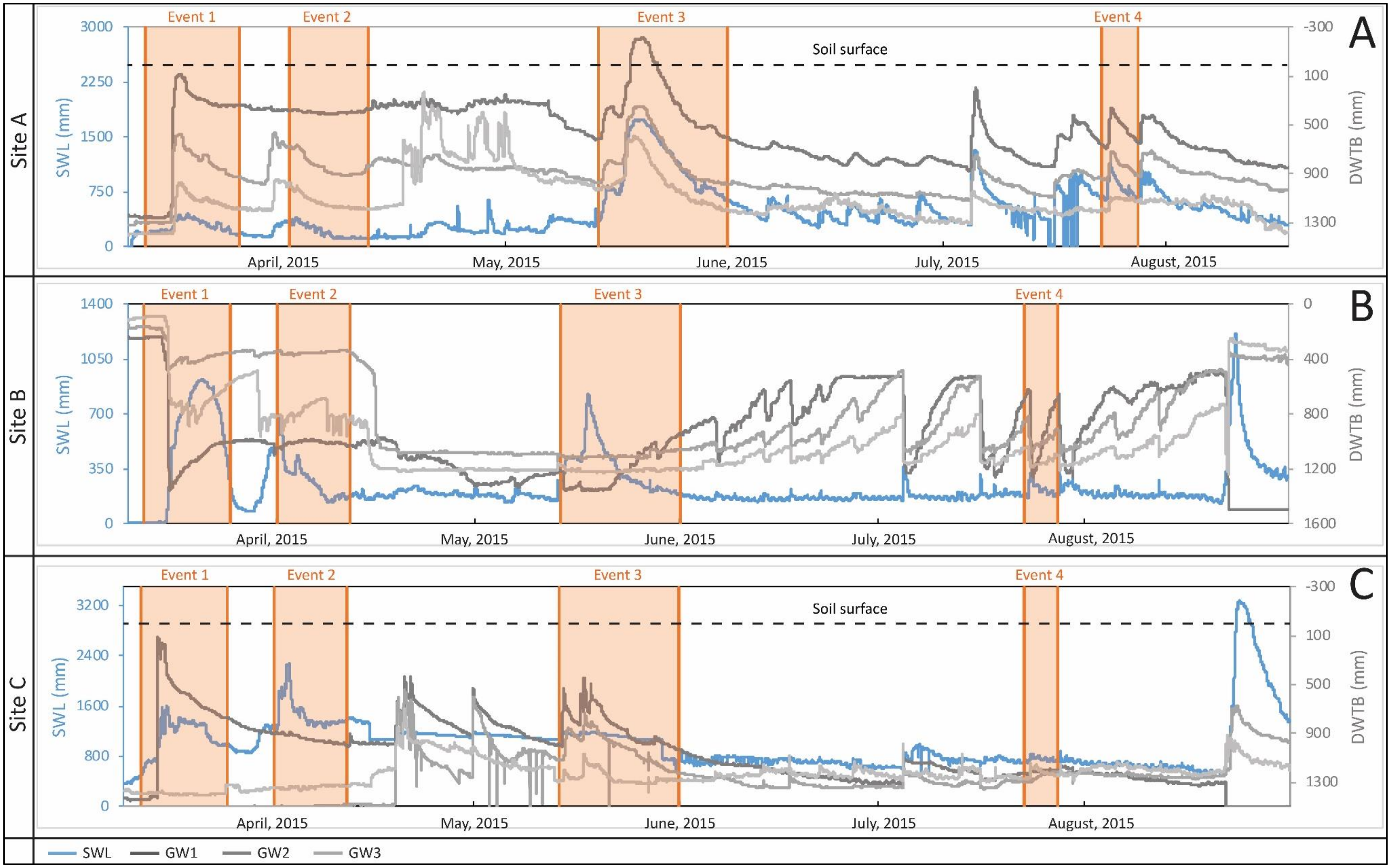

Figure 4.2. Timeseries of depth to water table (DWTB) in the studied riparian areas as well as water levels (SWL) in the adjacent drainage channels in spring and summer 2015 for sites A, B and C. GW1, GW2 and GW3 refer to the three groundwater wells in each riparian area. Orange shaded areas show the events targeted for further analysis in the current chapter.

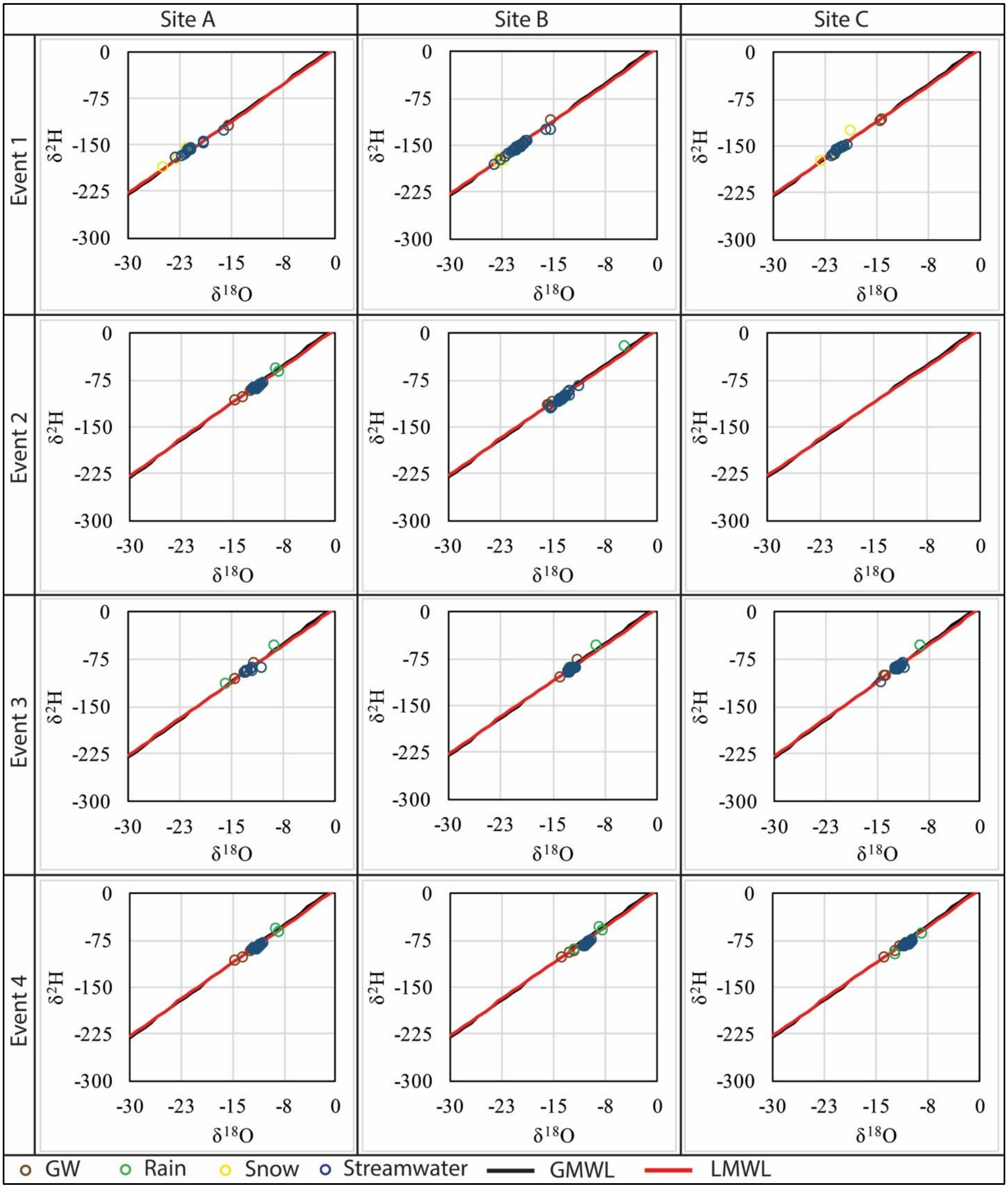

Figure 4.3. Event-specific groundwater (GW), precipitation (rain and snow) and streamwater isotopic signatures relative to the global meteoric water line (GMWL: $\delta^2H = 8 * \delta^{18}O + 10$) and the local meteoric water line (LMWL: $\delta^2H = 7.78 * \delta^{18}O + 6.22$).

Hydrograph separation showed that regardless of event type, a significant portion of streamflow, typically more than 25 %, was attributable to ‘old’ water sources (Tables 4.2 and 4.3). A statistical comparison of $PROP_{old}$ characteristics between snowmelt- and rainfall-triggered events using the Kruskal-Wallis test showed that the median $PROP_{old}$ was smaller and the maximum $PROP_{old}$ was larger for snowmelt-triggered events, but those differences were not statistically significant at the 95 % level (Figure 4). Similarly, the lags to $PROP_{old}$ rise and peak were longer for snowmelt-triggered events; however, those differences were only statistically significant for the lag to $PROP_{old}$ peak.

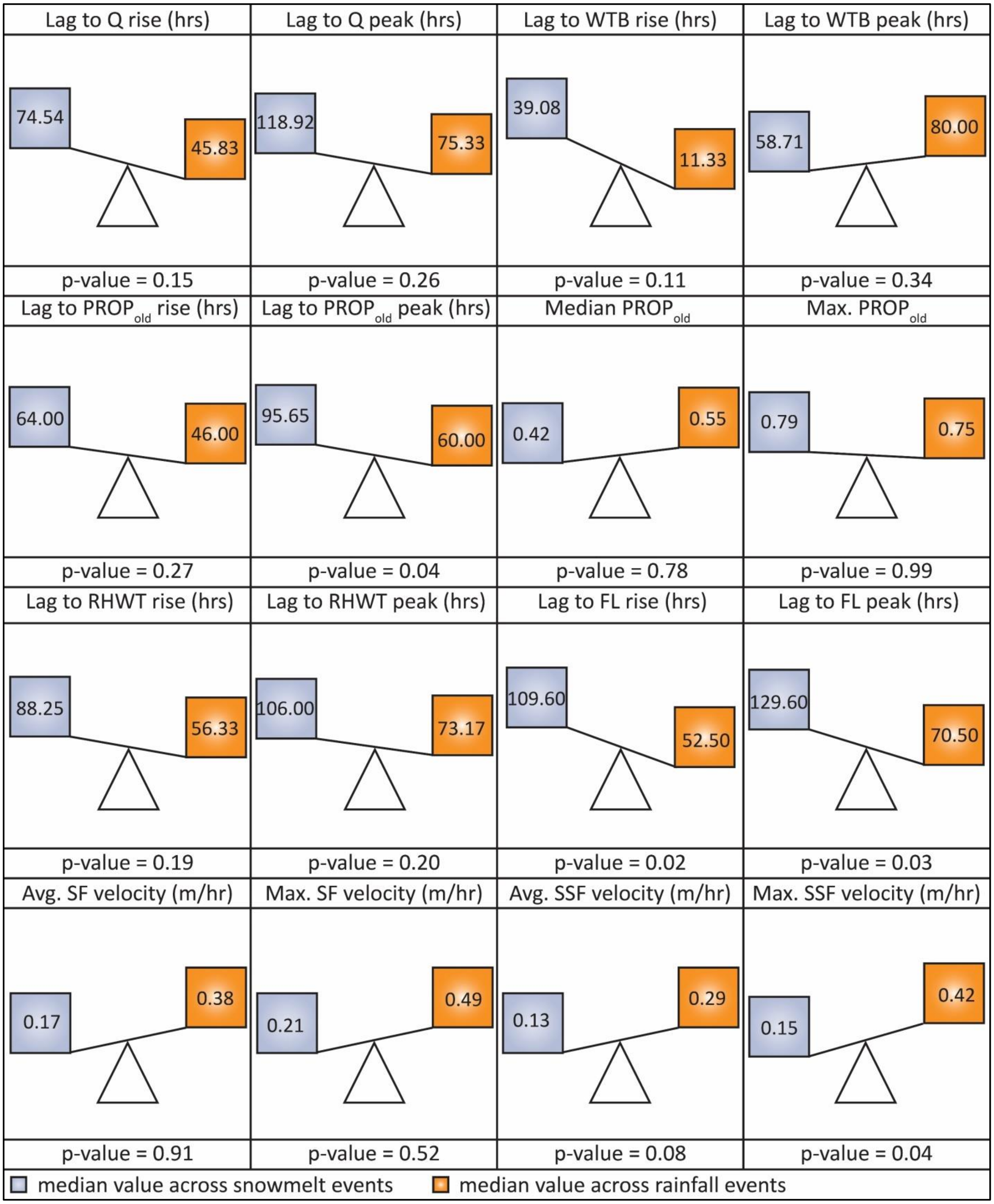

Figure 4.4. Comparison of selected snowmelt and rainfall event characteristics. Q: stream discharge, WTB: water table; $PROP_{old}$: proportion of old water in streamflow; RHWT: rhodamine; FL: fluorescein; SF: surface flow; SSF: subsurface flow; Avg.: average; Max.: maximum. p-values lesser than 0.05 mean that the null hypothesis associated with the Kruskal-Wallis test can be rejected, i.e., snowmelt and rainfall-triggered event characteristics are significantly different from one another.

For the first snowmelt-triggered event (Event 1), Sites A and B shared similar $PROP_{old}$ dynamics while Site C differed significantly (Table 4.2 and Figure 4.5). In general, fluctuations in $PROP_{old}$ at Sites A and B were synchronized with total flow (i.e., hydrograph) dynamics, with the exception of a few singular peaks in the $PROP_{old}$ timeseries at Site B. Site C had much lower $PROP_{old}$ than Sites A and B throughout Event 1, and several hydrograph peaks were not mirrored in the $PROP_{old}$ dynamics at that site. Lags to $PROP_{old}$ rise and peak were comparably large between Sites A and B, while the lag to $PROP_{old}$ rise at Site C was almost null and peak $PROP_{old}$ was reached more quickly. For Event 2, statistics related to $PROP_{old}$ at Site C are unavailable due to missing data; however, at Sites A and B, summary statistics for the $PROP_{old}$ and flow timeseries were nearly identical, except that the lags to $PROP_{old}$ rise and peak were longer at Site A than at Site B. Additionally, at Site B, DWTB fluctuations in GW3 later in the event were aligned with increases in $PROP_{old}$ that approached 1.0. In general, $PROP_{old}$ appeared to be greater during Event 2 compared to Event 1. It is also notable that air temperature during Event 2 was highly variable and multiple fluctuations above and below 0 degrees Celsius were observed.

Table 4.2. Precipitation, snow water equivalent (SWE), streamflow and water table dynamics, isotopic and fluorescent dye tracer characteristics associated with snowmelt-triggered events. For each variable, values reported left and right of the dash are for Event 1 and Event 2, respectively. "Temporal window" refers to the antecedent temporal window used for some computations. See text for definitions of SWE-related variables. "DWTB": depth to the water table; $PROP_{old}$: proportion of old water in streamflow; [ ]: concentration; Stdev: standard deviation. NA: data not available.

	Site A			Site B			Site C		
Antecedent precipitation characteristics									
Temporal Window (Days)	1	7	30	1	7	30	1	7	30
AP_{xday} (cm)	0.0-0.2	55.3-30.0	55.4-190	0.1-0.1	33.5-18.8	38.5-130	0.0-0.1	35.6-13.4	40.5-110
Total rainfall (mm)	0.3-1.0			0.4-2.2			0.4-2.3		
Snow water equivalent characteristics									
Available SWE (cm)	67.5-33.4			42.4-16.7			43.9-14.5		
7-day melted SWE (cm)	55.2-29.8			33.2-18.7			35.2-13.3		
Total event SWE (cm)	87.4-30.1			66.6-16.1			50.6-14.3		
Riparian water table and adjacent streamflow dynamics									
Peak Q (m^3/s)	0.3-0.3			0.2-0.1			264-55494		
Peak channel fullness (%)	32.4-19.1			92.5-41.1			>100->100		
Lag to hydrograph rise (hrs)	94.0-115			86.0-69.0			62.0-21.3		
Lag to hydrograph peak (hrs)	144-133			210-77.0			94.0-55.5		
Minimum DWTB (mm)	546-680			662-719			998-1241		
Lag to WTB rise (hrs)	69.5-0.0			61.8-13.3			61.3-28.8		
Lag to WTB peak (hrs)	113-1.0			96.8-17.5			62.8-61.8		
Max WTB increase (mm)	771-153			669-103			445-33.5		
Old water temporal dynamics									
Mean $PROP_{old}$	0.3-0.8			0.3-0.8			0.1-NA		
Median $PROP_{old}$	0.2-0.7			0.3-0.8			0.0-NA		
Minimum $PROP_{old}$	0.0-0.7			0.0-0.6			0.0-NA		
Maximum $PROP_{old}$	0.9-0.9			0.9-1.0			0.2-NA		
Stdev $PROP_{old}$	0.2-0.1			0.2-0.1			0.1-NA		
Lag to $PROP_{old}$ rise (hrs)	70.0-115			66.0-69.0			0.0-NA		
Lag to $PROP_{old}$ peak (hrs)	108-133			102-77.0			58.0-NA		
# of peaks in $PROP_{old}$	3.0-6.0			4.0-7.0			3.0-NA		

Dye-tracer breakthrough curve characteristics	Site A RHWT	Site A FL	Site B RHWT	Site B FL	Site C RHWT	Site C FL
Mean [ ] (ppb)	2.9-0.7	1.1-1.6	0.8-0.5	0.9-1.3	3.3-NA	1.3-NA
Median [ ] (ppb)	0.3-0.6	1.0-1.6	0.3-0.4	0.8-1.3	0.5-NA	1.2-NA
Minimum [ ] (ppb)	0.0-0.2	0.8-1.0	0.0-0.0	0.6-1.1	0.0-NA	0.9-NA
Maximum [ ] (ppb)	12.3-1.9	2.3-1.7	6.9-1.6	1.7-1.5	18.8-NA	1.9-NA
Stdev [ ] (ppb)	4.7-0.3	0.4-0.1	1.6-0.4	0.3-0.1	6.4-NA	0.3-NA
Lag to [ ] rise (hrs)	70.0-111	102-113	70.0-NA	70.0-161	102-NA	102-NA
Lag to [ ] max (hrs)	102-137	102-177	75.0-NA	90.0-169	110-NA	110-NA
# of peaks in [ ]	3.0-6.0	2.0-7.0	3.0-NA	3.0-4.0	4.0-NA	4.0-NA
Average travel velocity (m/hr)	0.2-0.2	0.2-0.1	0.1-NA	0.1-0.1	0.1-NA	0.1-NA
Maximum travel velocity (m/hr)	0.3-0.2	0.2-0.2	0.2-NA	0.2-0.1	0.2-NA	0.2-NA

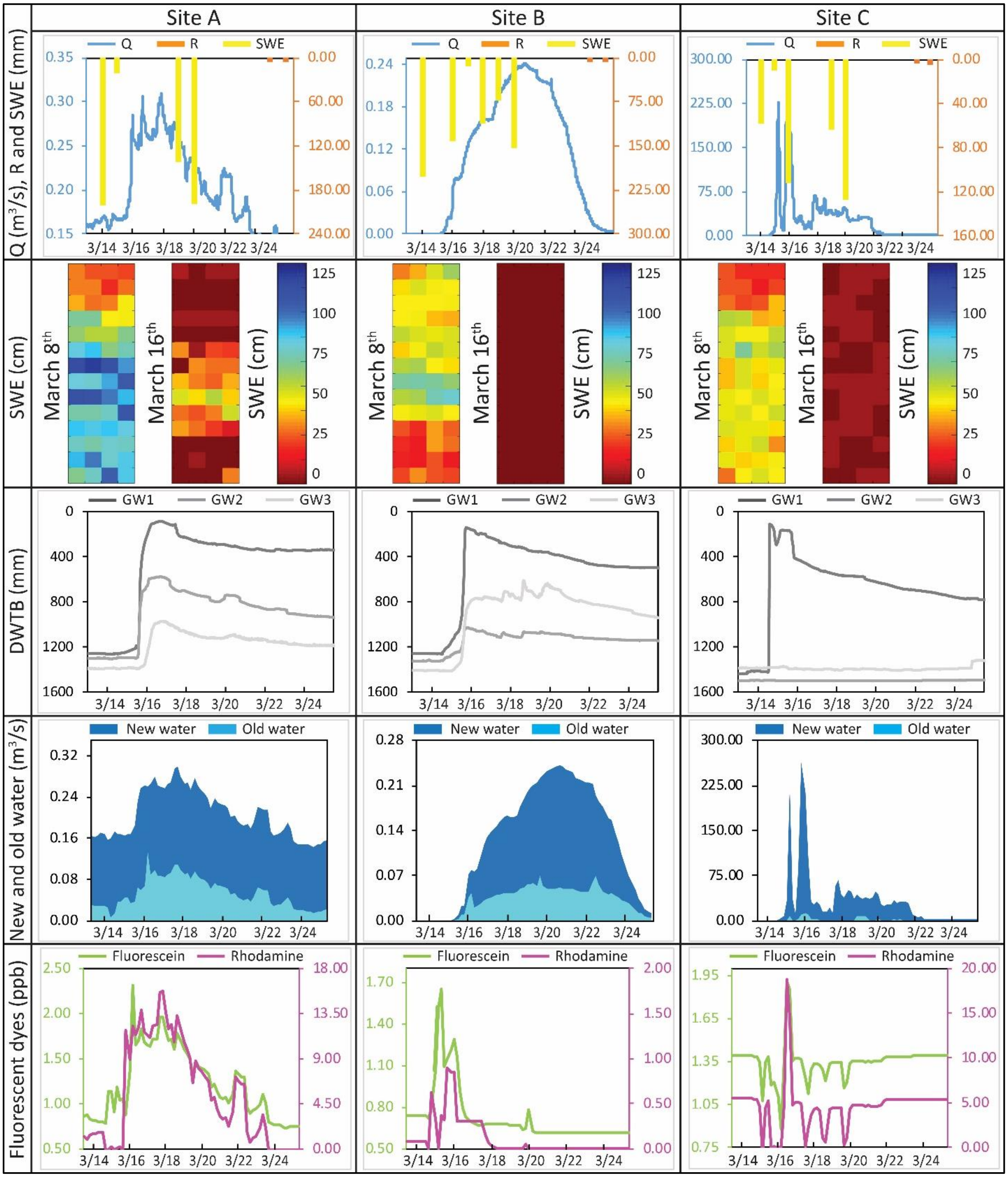

Figure 4.5. Discharge (Q), rainfall (R), melted snow expressed as snow water equivalent (SWE), depth to water table (DWTB) and fluorescent dye tracer breakthrough curves associated with Event 1. For SWE maps, the orientation of the experimental plot is the same as shown in Figure 4.1.

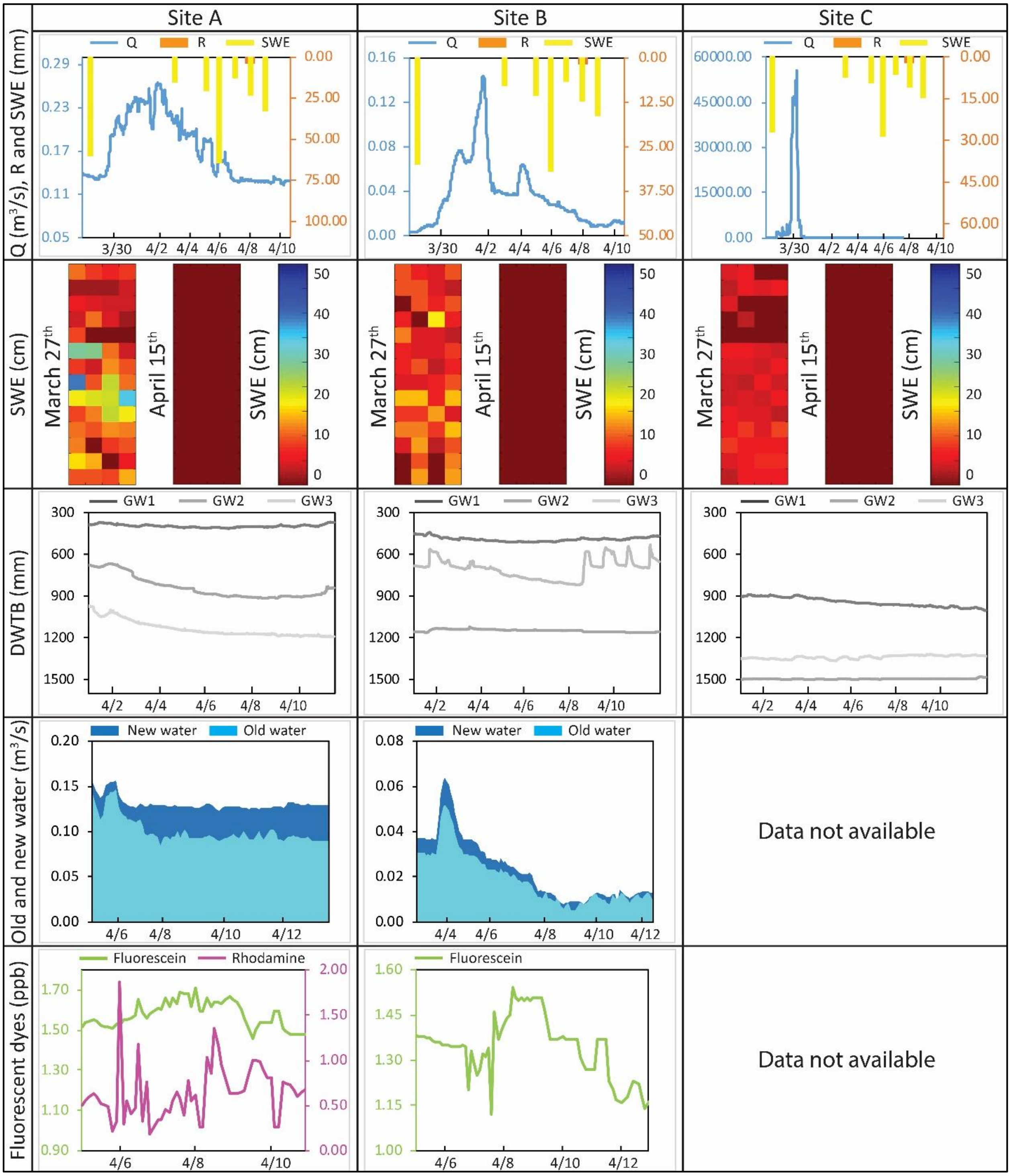

Figure 4.6. Discharge (Q), rainfall (R), melted snow expressed as snow water equivalent (SWE), depth to water table (DWTB) and fluorescent dye tracer breakthrough curves associated with Event 2. For SWE maps, the orientation of the experimental plot is the same as shown in Figure 4.1.

Regarding rainfall-triggered events, during Event 3, $PROP_{old}$ summary statistics were similar at all sites. $PROP_{old}$ values were relatively constant throughout the whole event as evidenced by their small range and standard deviation (Table 4.3 and Figure 4.7). The lags to $PROP_{old}$ rise and peak were longer at Site A, compared to Sites B and C, and at all sites only a short period of time separated initial $PROP_{old}$ response and peak conditions. Also, $PROP_{old}$ and hydrograph dynamics were synchronized at all sites; however, at Site C, small short-term fluctuations in $PROP_{old}$ were especially evident during the second half of the event (Figure 4.7). Overall, for Event 4, $PROP_{old}$ was small relative to other events and also did not vary much throughout the event. Event 4 was characterized by a large number of (and mostly synchronized) small peaks in the event hydrograph and the $PROP_{old,}$ timeseries. The lags to $PROP_{old}$ rise and peak were longest at Site A, followed by Site B and Site C but in absolute terms, they were significantly shorter than those observed with other events.

Table 4.3. Precipitation, soil moisture (SM), streamflow and water table dynamics, isotopic and fluorescent dye tracer characteristics associated with rainfall-triggered events. For each variable, values reported left and right of the dash are for Event 3 and Event 4, respectively. "Available SM" refers to the average SM in the riparian area preceding the event. "Event SM" refers to the average SM during the event. See the caption of Table 4.2 for abbreviations.

	Site A			Site B			Site C		
Antecedent rainfall characteristics									
Temporal Window (Days)	1	7	30	1	7	30	1	7	30
AP_{xday} (cm)	15.4-11.9	16.5-25.3	32.5-64.6	13.6-4.7	14.3-11.9	38.0-63.8	13.4-4.0	14.1-10.7	38.5-63.7
Total rainfall (mm)		41.8-31.1			47.6-25.9			48.10-25.4	
Snow water equivalent characteristics									
Available SM (%)		85.6-80.9			96.6-96.1			86.9-86.4	
Event SM (%)		86.2-81.8			98.2-77.4			92.9-93.6	
Riparian water table and adjacent streamflow dynamics									
Peak Q (m^3/s)		10.2-2.0			0.2-0.1			10.8-1.1	
Peak channel fullness (%)		>100-83.5			>100-54.0			>100->100	
Lag to hydrograph rise (hrs)		73.5-18.5			73.5-18.0			74.0-17.5	
Lag to hydrograph peak (hrs)		146-30.8			98.0-31.8			15.5-130	
Minimum DWTB (mm)		259-730			265-332			760-1177	
Lag to WTB increase (hrs)		9.0-18.0			8.8-16.8			10.8-3.5	
Lag to WTB peak (hrs)		120-31.8			75.0-32.8			91.3-130	
Max WTB increase (mm)		610-208			70.9-296			372-88.6	
Old water temporal dynamics									
Mean $PROP_{old}$		0.7-0.5			0.7-0.4			0.6-0.4	
Median $PROP_{old}$		0.7-0.5			0.7-0.4			0.6-0.4	
Minimum $PROP_{old}$		0.3-0.4			0.6-0.3			0.4-0.2	
Maximum $PROP_{old}$		0.8-0.6			0.8-0.5			1.0-0.9	
Stdev $PROP_{old}$		0.1-0.1			0.1-0.0			0.1-0.1	
Lag to rise $PROP_{old}$ (hrs)		101-37.0			50.0-28.0			44.0-16.0	
Lag to peak $PROP_{old}$ (hrs)		101-67.0			59.0-46.0			50.0-37.0	
# of peaks in $PROP_{old}$		1.0-7.0			1.0-9.0			5.0-7.0	

Dye-tracer breakthrough curve characteristics	Site A RHWT	Site A FL	Site B RHWT	Site B FL	Site C RHWT	Site C FL
Mean [ ] (ppb)	2.0-1.5	2.0-2.4	1.2-2.4	1.6-2.4	2.2-2.1	1.3-1.8
Median [ ] (ppb)	1.4-1.4	2.0-2.4	0.8-3.1	1.5-2.3	0.7-2.1	1.2-1.7
Minimum [ ] (ppb)	0.8-1.0	1.8-2.2	0.0-1.1	1.4-2.1	0.2-0.8	1.1-1.4
Maximum [ ] (ppb)	8.8-3.3	2.3-2.9	5.2-3.6	2.0-4.9	9.2-3.8	1.8-4.8
Stdev [ ] (ppb)	1.7-0.5	0.1-0.2	1.2-1.0	0.2-0.4	2.6-1.0	0.2-0.5
Lag to [ ] rise (hrs)	77.0-34.0	86.0-28.0	83.0-55.0	68.0-13.0	89.0-0.0	77.0-43.0
Lag to [ ] max (hrs)	101-40.0	101-61.0	93.0-79.0	93.0-16.0	113-13.0	107-61.0
# of peaks in [ ]	3.0-6.0	3.0-4.0	6.0-4.0	5.0-1.0	3.0-4.0	4.0-1.0
Average travel velocity (m/hr)	0.2-0.6	0.2-0.4	0.1-0.1	0.1-0.6	0.1-1.1	0.1-0.2
Maximum travel velocity (m/hr)	0.3-0.7	0.3-0.8	0.1-0.2	0.2-0.8	0.2-1.5	0.2-0.3

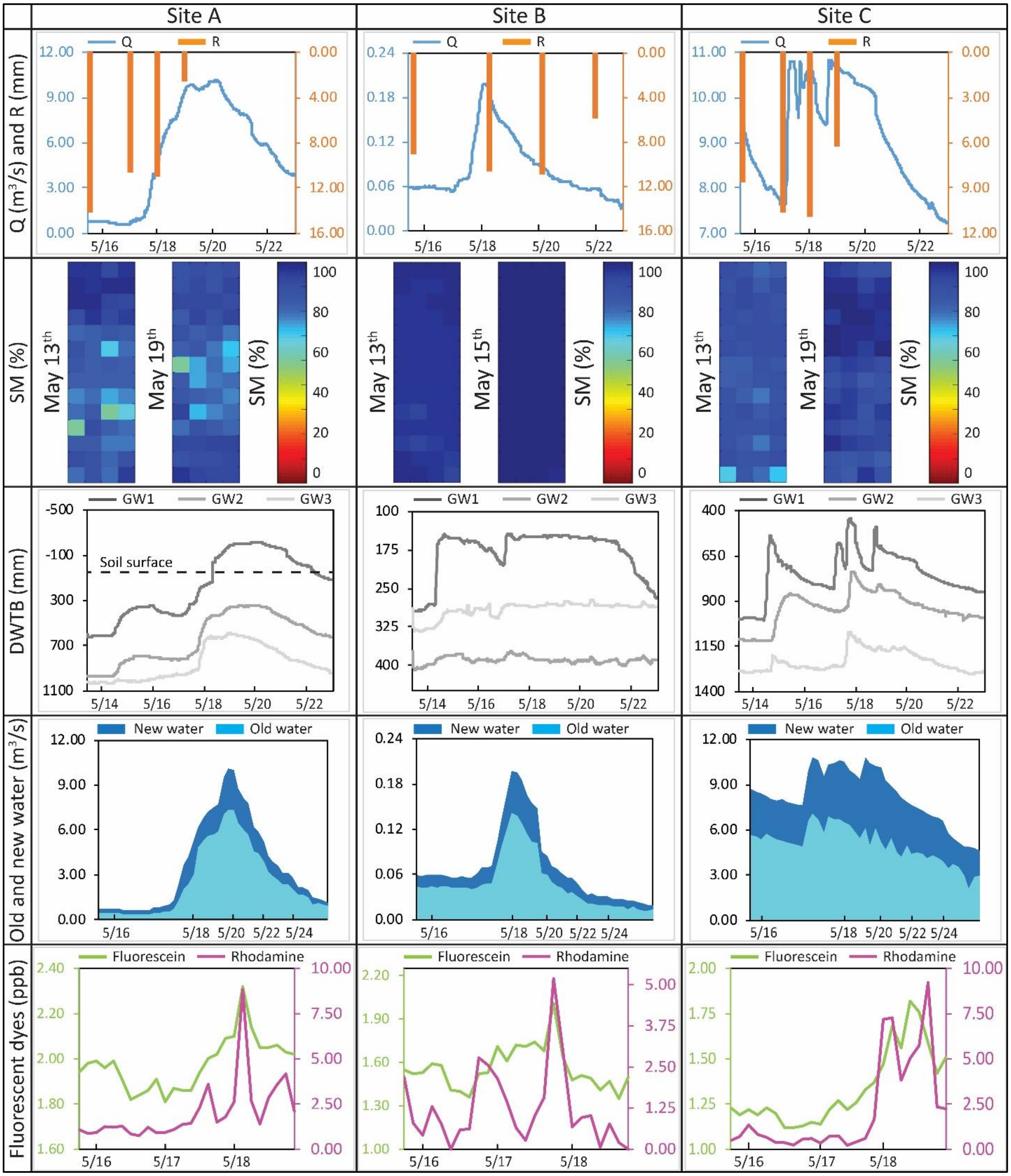

Figure 4.7. Discharge (Q), rainfall (R), soil moisture (SM), depth to water table (DWTB) and fluorescent dye tracer breakthrough curves associated with Event 3. For SM maps, the orientation of the experimental plot is the same as shown in Figure 4.1.

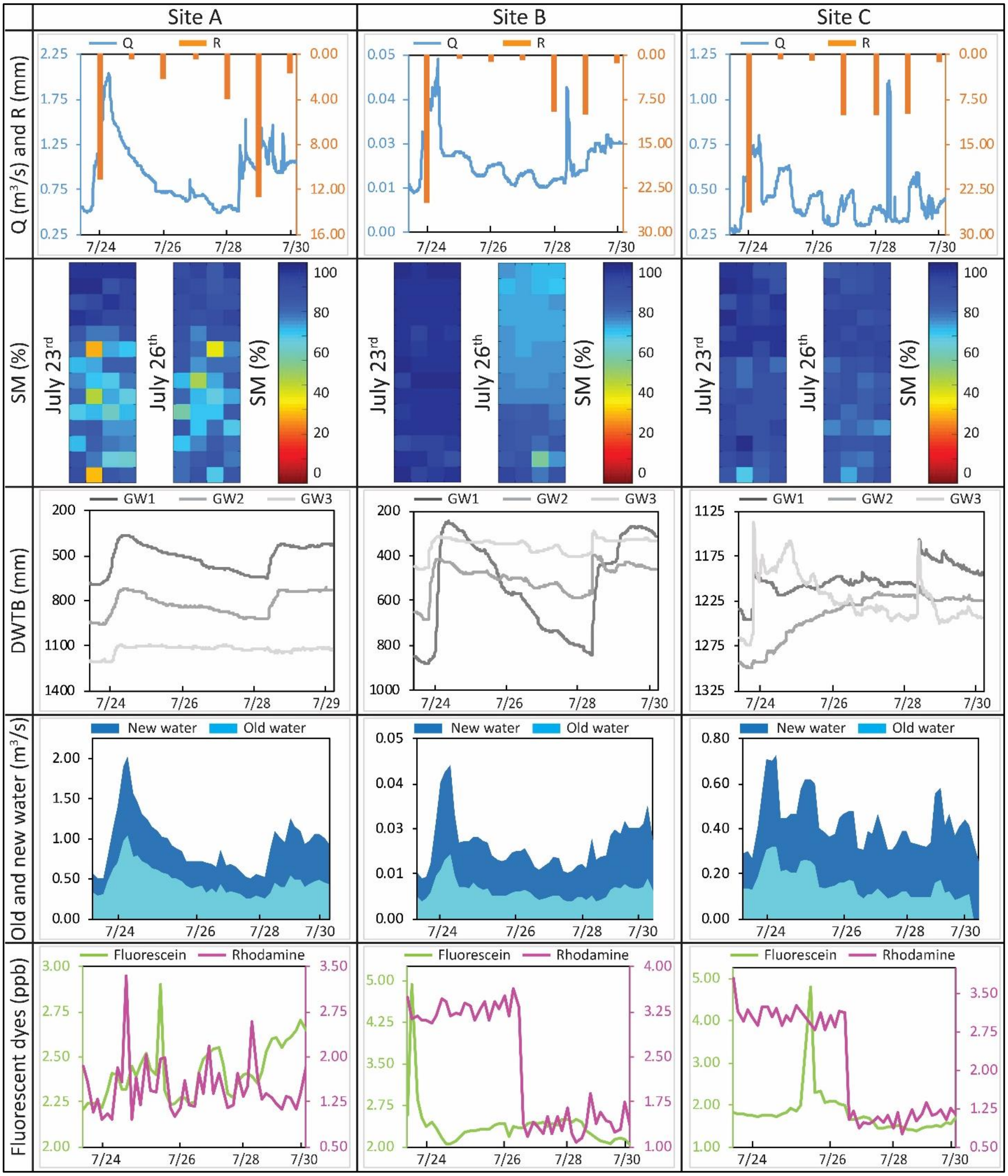

Figure 4.8. Discharge (Q), rainfall (R), soil moisture (SM), depth to water table (DWTB) and fluorescent dye tracer breakthrough curves associated with Event 4. For SM maps, the orientation of the experimental plot is the same as show in Figure 4.1.

4.3.2 - Influence of snow cover extent and AMCs on event response characteristics and temporal and geographic sources of streamflow

For Event 1, SWE characteristics (i.e., available SWE, 7-day melted SWE and total event SWE) were similar at Sites B and C, but greater SWE amounts were observed at Site A (Table 4.2). Riparian SWE maps showed heterogeneous snowpack distributions prior to Event 1 and nearly complete snowpack depletion throughout the event (Figure 4.5). Water inputs associated with melting snow initiated the hydrograph rising limb at all sites. At Site B, flow recession progressed without further inputs, while Sites A and C had new snowmelt and rain during flow recession that did not initiate new hydrologic response. In general, while event hydrograph rise and recession at Sites A and B occurred progressively over a relatively long period of time, response was 'flashy' and occurred at a larger magnitude at Site C. In terms of WTB, the maximum event rise was achieved shortly after initial response and followed by a gradual recession in WTB at all sites (Figure 4.5). Those dynamics were observed in each groundwater well, with the exception of GW2 and GW3 at Site C where subsurface water remained frozen throughout the event. At Sites A and C, the shapes of dye breakthrough curves for RHWT and FL were similar to that of the hydrographs. At Site B, however, both breakthrough curves were synchronized but RHWT and FL concentrations peaked and returned to background concentrations rapidly, a flashy dye tracer response which deviated from the wider, bell-shaped curves associated with the hydrograph and the $PROP_{old}$ timeseries.

Comparatively for Event 2, SWE maps indicated a heterogeneous snowpack distribution at Sites A and B, but a more uniform distribution at Site C prior to the event and complete snowpack depletion within the event duration at all sites (Figure 4.6). At all sites, melting snow initiated the rising limb of the hydrograph, and inputs later in the event during hydrograph flow recession did not result in the initiation of additional hydrologic response. Similar to Event 1, at Site A and B, hydrograph rise and recession occurred progressively over a relatively long period of time, while response was 'flashy' and occurred at a larger magnitude at Site C. Water inputs did not result in WTB response during Event 2, except for Site B where a straightforward precipitation-WTB response in GW3 was observed. Dye breakthrough curves for FL that were available for Event 2 (i.e., Sites A and B) resembled the general shape of event hydrographs; however, breakthrough characteristics, like the initial rise in dye concentration and peak, were slightly delayed relative to the lag to hydrograph rise and peak. In contrast, RHWT breakthrough curves for Site A showed rapid fluctuations of in-stream RHWT concentrations throughout Event 2 and initial increase and peak concentration occurred prior to FL.

The ranking of events according to decreasing AMCs for a 30-day antecedent window yielded the following order: Event 2, Event 4, Event 1 and Event 3. Relatively slow and progressive hydrograph rise and recession were observed at Site A for both high AMCs (Event 2) and low AMCs (Event 3) while at Sites B and C, hydrograph response for high AMCs was more ‘flashy’ with narrower peaks and especially high Q magnitudes at Site C. However, hydrograph characteristics for events with intermediate AMCs showed greater variability in hydrograph shape and had slightly faster hydrograph rise and recession in some cases at Sites A and B and decreased Q magnitude with less ‘flashy’ dynamics at Site C. For high and low AMCs, additional water inputs following the initial hydrograph rise and peak did not result in additional response. In contrast, for intermediate AMCs, additional water inputs during hydrograph recessions led to additional, although smaller-magnitude hydrograph rises. Events associated with intermediate AMCs showed numerous similarities in response timing and magnitude between $PROP_{old}$ and hydrograph dynamics, and between DWTB and both hydrograph dynamics and precipitation. Comparatively, high and low AMC-events yielded more pronounced maximum conditions in $PROP_{old}$ than intermediate AMC-events. Often, regardless of high, low or intermediate AMCs, fluctuations in RHWT followed shortly after precipitation. Fluctuations in FL breakthrough curves came shortly after changes in DWTB and often breakthrough curves were similar, in shape, to DWTB timeseries.

4.3.3 - Travel patterns of surface and subsurface water

The travel velocities of surface water labelled with RHWT were relatively similar for Events 1, 2 and 3 (Tables 4.2 and 4.3). In general, for those events, average and maximum RHWT travel velocities were approximately 0.2 and 0.3 m/hr, respectively, with the exception of missing data for Sites B and C during Event 2. For Event 4, the average and maximum RHWT travel velocities at Sites A and C were 2.5 to 11 times larger than those observed for other events, while travel velocities at Site B remained similar to preceding events.

When considering subsurface travel velocities, quantified using FL dye, average and maximum travel velocities were also relatively similar for Events 1, 2 and 3. Average FL travel velocities ranged between 0.1 and 0.2 m/hr, while maximum FL travel velocities ranged between 0.1 and 0.3 m/hr. For Event 4, however, significant increases in average and maximum FL travel velocities were observed at Sites A and B but not at Site C. In assessing subsurface flow direction, relatively low gradients were observed for all sites, with the exception of a considerable increase in gradient magnitude during Event 4 (Table 4.4). Additionally, at Sites A and C, the hydraulic gradient was typically positive, while at Site B gradients were usually negative, with the exception of Event 4 (Table 4.4). Comparison of subsurface and surface water travel velocities based on event type using the Kruskal-Wallis test showed that the average and maximum travel velocities were faster for rainfall-triggered events, but the differences between event types were only statistically significant when considering the maximum subsurface velocity (Figure 4.4).

Table 4.4. Statistical summary of event-specific riparian hydraulic gradient characteristics. Stdev: standard deviation; CV: coefficient of variation.

Hydraulic gradient characteristics												
	Event 1			Event 2			Event 3			Event 4		
Statistic	Site A	Site B	Site C	Site A	Site B	Site C	Site A	Site B	Site C	Site A	Site B	Site C
Mean	0.02	-0.02	0.03	0.02	-0.01	0.04	0.02	-0.00	0.04	0.03	0.02	0.07
Median	0.01	-0.02	0.02	0.02	-0.01	0.04	0.02	-0.00	0.04	0.02	0.02	0.07
Minimum	-0.01	-0.05	-0.02	0.01	-0.02	0.04	0.00	-0.01	0.02	0.02	0.00	0.07
Maximum	0.06	-0.00	0.07	0.03	0.00	0.05	0.04	0.01	0.05	0.03	0.04	0.08
Stdev	0.02	0.01	0.02	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.01	0.00
CV	1.03	-0.43	0.84	0.22	-0.52	0.06	0.57	-1.69	0.16	0.22	0.69	0.03

4.4 Discussion

4.4.1 - Streamflow temporal sources and the influence of event type on their relative importance

To date, research on runoff generation mechanisms in the Prairies has mainly focused on surface runoff processes since snowmelt runoff, often occurring over frozen ground, was believed to represent 80 % of the total annual Prairie runoff (Gray and Landine, 1988). However recently, due to climate and land-use changes, increases in the rainfall fraction of total precipitation and increases in the frequency of multi-day rainfall events have been observed in the Prairies (Dumanski et al., 2015), thus creating a need for a better understanding of non-snow-dominated Prairie hydrology.

Studies in regions other than the Prairies have found significant streamflow contributions from ‘old’ water. For example, riparian groundwater accounted for 35 to 50 % of storm runoff during two winter rainstorms at the Panola Mountain Research Watershed (Georgia, USA, Burns et al., 2001), while 25 to 75 % of storm runoff was attributable to shallow subsurface flow during a summer storm event at the Shelter Creek Watershed (New York, USA; Brown et al., 1999). Previous studies have also found that stormflow is dominated (> 60 %) by ‘old’ water in small forested or wetland headwater watersheds in Canada (Gibson et al., 2005). Results from the CCW were in partial agreement with those findings as all of our monitored events showed significant ‘old’ water contributions (average between 30 and 80 %) to streamflow (Tables 4.2 and 4.3). The relatively high importance of ‘old’ water, even during snowmelt-triggered events, is therefore a deviation from traditional conceptualizations of Prairie runoff generation that consider the overland flow of new water to be dominant. It is also not in agreement with research from arctic environments, despite the fact that both the CCW and Arctic environments share some common characteristics related to snowmelt dynamics over frozen ground: arctic streamflow during snowmelt was found to be primarily ‘new’ water (McNamara et al., 1997). Seasonal variability in ‘old’ water contributions has been previously observed (e.g., McNamara et al., 1997), which is aligned with results from the CCW where ‘old’ water contributions during snowmelt-triggered events were smaller than during rainfall-triggered events, thus raising the question of how ‘old’ water is mobilized quickly during events. The rapid mobilization of ‘old’ water to streamflow has received significant attention in the literature (e.g., Kirchner, 2003) and has been linked to several potential mechanisms. For example, at a steep, humid watershed in New Zealand, it was hypothesized that rapidly infiltrated water to the bedrock via macropores mixes with ‘old’ water sitting in bedrock hollows and causes a rapid movement of ‘old’ water into the stream during storm events via piston flow (McDonnell, 1990). Alternatively, the transmissivity feedback hypothesis suggests that during snowmelt, water flux increases along with the water table into soil horizons with high saturated hydraulic conductivity, resulting in ‘old’ water mobilization and rapid travel to the stream through shallow flowpaths (Kendall et al., 1999). The transmissivity feedback hypothesis is plausible for the sites we monitored in the CCW, especially during Event 2 as we observed an increase in ‘old’ water contributions to the stream and decreased DWTB values (especially in GW1, i.e., in near-stream areas). However, while transmissivity feedback might explain ‘old’ water mobilization during snowmelt-triggered events, it does not necessarily explain observations from rainfall-triggered events. Close proximity of the water table to the ground surface not only implies that ‘old’ water will be able to access shallow subsurface pathways (Brady et al., 1996), but also that only a small amount of water from precipitation is required to saturate the soil (Buttle, 1994) and return flow can occur. During Event 3, in particular, we observed low or null DWTB and relatively high $PROP_{old}$, suggesting the occurrence of return flow (Figure 4.7). Additionally, similar timing in FL and RHWT breakthrough curves during Event 3 may suggest mixing between ‘old’ and ‘new’ water and hence indicate the simultaneous occurrence of saturation-excess overland flow – from rain falling on saturated soils – and return flow (Dunne and Black, 1970).

4.4.2 - Influence of AMCs and snow cover on event response and geographic sources of streamflow

It is well known that precipitation intensity and duration influence event hydrograph characteristics (i.e., magnitude and timing of response and peak) (Davie, 2008). In cold weather regions, in particular, snowpack depth has been found to affect response ‘flashiness’: shallow snowpacks can produce quick runoff over frozen ground but lower intensity and longer duration water inputs in the form of melt water may result in a wider hydrograph peak and a longer recession limb (Quick and Pipes, 1976). Results from the CCW met those expectations: the reduction in snow cover extent during Event 2 was accompanied by elevated Q and narrower hydrograph peaks at Site C, relative to Event 1 (Figures 4.5 and 4.6). Snowmelt initiation and rapid streamflow fluctuations have also been associated with changes in temperature surrounding 0 degrees Celsius (Jin et al., 2012). Similarly, at Sites B and C in the CCW, multiple fluctuations in temperature surrounding 0 degrees Celsius during Event 2 yielded relatively narrow and more pronounced hydrograph peaks (Figure 4.6).

AMCs have been shown not only to influence stream response to precipitation but also streamflow composition during a single storm event (Cey et al., 1998). At the Kin-tore Creek watershed (Southern Ontario, Canada), larger increases in stream Q were observed after wet antecedent conditions (Cey et al., 1998). Similarly, for the CCW, we observed larger increases in stream Q during Event 3 as opposed to Event 4, likely due to wetter antecedent conditions (Figures 4.7 and 4.8). In general, elevated AMCs have been shown to result in larger total and peak storm Q (Sklash et al., 1986; Hinton et al., 1994; Dingman, 2015). However, in the CCW, elevated AMCs only resulted in larger peak Q in some cases. For example, the largest Q values at Site C were coincident with the high AP_7 and AP_{30} values of Event 2 while at Sites A and B, the largest Q values were associated with Event 3. Differences in the results observed at the CCW are likely the result of a combination of additional variables, including event type and precipitation rate and intensity.

4.4.3 - Travel patterns of subsurface and surface water to streams

In terms of subsurface flow direction, flow reversals in the CCW were observed at Site B during Events 1, 2 and 3 and were associated with decreases (rises) in DWTB (WTB) (Table 4.4). This is in agreement with research conducted in Indiana where a reduction in hydraulic gradient initiated by a rapid rise in the water table resulted in flow reversals (Vidon, 2012). The cause of flow reversals at Site B in the CCW is likely the design of downstream culverts; those culverts are intentionally undersized and slightly elevated above the channel bed to control channel velocity and downstream water movement. This design may result in artificially high channel levels that are at a higher elevation than the water table, subsequently leading to flow reversals.

The subsurface flow velocities estimated in the current study were significantly different than those reported in the literature. For instance, in forested catchments in New Zealand, fast subsurface flow velocities (i.e., average and maximum of 10.8 and 15.12 m/hr respectively) resulted in significant subsurface contributions to streamflow (Mosley, 1979). Subsurface flow velocities in the CCW were significantly lower, and the larger topographic relief present in the New Zealand watersheds was likely the cause of faster subsurface flow velocities in that area. In the Trail Valley Creek research basin (Northwest Territories, Canada), that also has significant precipitation from snowmelt and frozen near-surface soil layers during the winter and spring, average shallow subsurface flow velocities between 0.04 and 4.0 m/hr were observed during snowmelt (Quinton and Marsh, 1998), compared to average subsurface flow velocities between 0.1 and 0.2 m/hr during snowmelt-triggered events in the CCW. Fast subsurface flow velocities at Trail Valley Creek were attributed to water table levels within the highly conductive near-surface peat layers but while the CCW has similar peat layers, their conductivity is likely different due to historical peatland drainage and changes in land use. Exceptionally, during Event 4, subsurface flow velocities exceeded surface flow velocities at Sites A and B. Such high subsurface flow velocities cannot be accounted for by saturation-excess overland flow or return flow alone and are likely a result of macropore flow, which consists of rapid vertical or lateral subsurface water movement through non-capillary cracks or channels (Beven and Germann, 1982). Such a scenario was likely in the CCW during Event 4: decreased DWTB at the beginning of the event likely allowed ‘old’ water to travel through near-surface macropores, such as cracks that commonly develop as soils dry during the summer months. Also, during Event 4, SM conditions were lower and the WTB did not rise as much as during Event 3, implying that unsaturated flow through shallow pathways was possible. The low subsurface flow velocities observed in the CCW, especially for Events 1, 2 and 3, may be partially attributed to the use of FL dye to evaluate subsurface flow velocities in this study. Since dye was injected into groundwater wells, early-travelling FL dye may have been slowed or lost via soil adsorption shortly after its introduction in the soil profile. While FL is known to have limited adsorption-related losses (Smart and Laidlaw, 1977), soil organic matter content can influence dye adsorption rate and magnitude. Besides, dye adsorption is known to be nonlinear depending on pore water flow velocities, with higher adsorption rates being associated with slower velocities and vice versa (Sabatini and Austin, 1991). In the CCW, soils usually have a non-negligible organic matter content and the observed dye breakthrough characteristics may have been exacerbated by slow pore water flow velocities. However, it is worth noting that the site with the largest soil organic matter content, i.e., Site, B, was the one associated with the largest average and maximum subsurface travel velocities during Event 4 (Table 4.3), thus suggesting that high organic matter does not necessarily lead to higher dye adsorption rates. Further, adsorption is also dependent on the type of subsurface flow mechanisms that are active: some mechanisms like preferential flow reduce the water contact time with adsorption sites, compared to mechanisms like piston flow for example (Feng et al., 2004). While process inferences appeared to be consistent between dye-tracing data and isotopic data, the impossibility to quantify dye adsorption losses is a limitation of the current study and might have led to a biased estimation of subsurface flow velocities using fluorescent dyes only.

4.5 Conclusion

In this study, streamflow sources and the influence of snow cover extent and AMCs on streamflow response characteristics were examined for snowmelt- and rainfall-triggered events in an eastern Prairie watershed. Contrary to traditional conceptualizations of Prairie runoff generation that consider overland flow to be dominant, this study showed significant ‘old’ water contributions for both snowmelt- and rainfall-triggered events. Those ‘old’ water contributions were seasonally variable, with larger contributions observed during rainfall-triggered events. Decreases in snow cover extent were accompanied by ‘flashy’ high magnitude streamflow response and in some cases, elevated antecedent moisture conditions resulted in larger peak discharge. Surface flow velocities were generally faster than subsurface flow velocities, with the exception of a summer rainfall event that was likely associated with increased ‘old’ water movement through near-surface macropores. Flow reversals were also observed under specific conditions, likely as a result of local culvert design leading to artificially high channel water levels that produced negative hydraulic gradients. The major contribution of the current study was therefore the demonstration that subsurface runoff generation can be significant year-round in Prairie landscapes, and this result has important implications for hydrological modelling. Additional field investigations are however needed to better understand the conditions that may promote the rapid delivery of ‘old’ water to streams in unique Prairie environments.

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CHAPTER 5.

FINAL SYNTHESIS AND CONCLUSIONS

5.1 Summary of conclusions and research significance

Prairie landscapes have received limited attention in the available hydrology literature, and interpretations pertaining to runoff generation for this region have traditionally been made through the modification of typical hillslope conceptualizations. In addition, currently available Prairie-focused hydrological literature tends to focus on dynamics prevailing during the spring snowmelt as well as on a large section of the central-to-western Prairies known as the Prairie Pothole Region. Such a season-specific and region (or landscape-specific) focus led to the assumption that Hortonian overland flow and or fill-and-spill processes dominate in Prairie landscapes. The presence of a mosaic of vegetation and soil conditions in the eastern Prairies however make it possible for shallow subsurface flow to play a major role in runoff generation via moderately to highly rapid subsurface flow pathways such as matrix flow, macropore flow and pipeflow. Since those subsurface pathways and their control factors have seldom been examined in the Prairies, the overall objective of this thesis was to address Prairie subsurface-flow-related knowledge gaps. Analysis of hydrologic data collected from a watershed located in south-eastern Manitoba, Canada, was achieved to examine the extent to which different mechanisms and factors can contribute to runoff and streamflow across the Prairies.

First, patterns of riparian hydrologic state variables (i.e., SM, SEC and ST) were analyzed to establish whether they could be used to infer soil water movement characteristics in the Prairies, in the same way that SM patterns have been used in more typical, temperate humid hillslope environments (Grayson *et al.*, 1997; James and Roulet, 2009; Ali and Roy, 2010). Specifically, riparian state variable characteristics were evaluated for their effectiveness as predictors of streamflow response, and characteristics of state variables were examined for switching behaviour between preferred states. Ultimately, the usefulness of riparian SM, SEC and ST for predicting streamflow response was not fully confirmed. However, potential evaporation, depth to water table and antecedent precipitation were found to influence riparian state variable characteristics significantly. Switching behaviour between dry and wet states in SM were observed, but were not confirmed for either SEC or ST. SM and SEC patterns were especially useful to infer vertical or lateral groundwater movements. Combined, these results suggest that spatial and non-spatial characteristics of riparian state variables can be useful for inferring riparian-to-stream water movement in near-level eastern Prairie landscapes like the Catfish Creek Watershed.

Second, hydrometric and tracer data were examined to determine whether shallow subsurface flow is a significant runoff generation mechanism in near-level eastern Prairie landscapes. Tracer data were notably used to identify the main geographic and temporal sources of streamflow, while the effects of event type (i.e., snowmelt- or rainfall-triggered) and AMCs on streamflow characteristics and sources were considered. Additionally, surface and subsurface flow characteristics were evaluated to make inferences about runoff-generation mechanisms. Results from this analysis confirmed that significant portions of streamflow are attributable to 'old' water regardless of event type and that AMCs do in fact influence streamflow characteristics. Further, surface flow velocities tended to be faster than subsurface flow velocities, with the exception of one event. It was hypothesized that the activation of macropore flow in some circumstances could have led to fast subsurface velocities. Those results have two major implications that go beyond the scope of the current thesis. First, the importance of subsurface flow is a critical component that needs to be captured and included in hydrological models used for flood forecasting, especially when subsurface flow velocities are so high that they may exceed surface velocities. Second, the large role of 'old' water in runoff generation means that contaminants present in the subsurface have a high likelihood of being mobilized during hydrological events and transported towards drainage channels. This is especially critical in the context of Prairie landscapes where a long history of agricultural activity and fertilizer inputs exists and has led to build-ups of nutrients in soils. Hydrological and biogeochemical models for the region therefore need to explicitly consider the role of 'old' water in nutrient export to better assess consequences for downstream water quality.

5.2 Remaining questions and opportunities for future research

Results obtained through this thesis work raise a number of questions in relation to the current project and future research. For questions pertaining to this project, one may consider how unmeasured upstream conditions might influence confidence in results. This particular concern does not apply to the headwater Site B and upstream conditions at engineered Sites A and C are consistent with the selected riparian sites and are not expected to introduce significant uncertainty in the presented results. Secondly, results in this thesis were presented with reference to both stream water level and stream discharge data, since depending on circumstance one type of data made it easier to highlight prevailing dynamics. This approach was merely intended to enhance visual clarity for interpretation and does not imply error or uncertainty in the rating curves developed to establish discharge from water levels. Finally, interpretations of breakthrough curves were completed in a qualitative fashion rather than using a 1-D dispersion model, despite having high-frequency fluorescent dye data: this approach was selected to avoid likely errors in identifying a dispersion coefficient for areas with highly variable soil characteristics.

The most evident questions relevant for future research are: (i) whether significant contributions to streamflow from subsurface sources are common in near-level Prairie landscapes; and (ii) which conditions promote the rapid delivery of ‘old’ water to streams in near-level Prairie landscapes? The significant variability present among Prairie landscapes makes answering those questions challenging, since organizing rules and principles are difficult to infer from high-frequency data only collected for a handful of sites and hydrologic events. The findings of the current study – and the impossibility to draw any generalization – highlight the fact that near-level Prairie landscapes, particularly unique areas like eastern Manitoba, are significantly under studied. Further investigations are therefore required to improve the overall understanding of subsurface flow in Prairie hydrology. To identify significant subsurface contributions to streamflow for the Prairie region in general, hydrologic studies comprising tracing experiments similar to those presented in this thesis (Chapter 4) could be undertaken for a larger variety of Prairie watersheds and events to cumulatively represent the physical and climatic heterogeneity characteristic of the region. Additionally, in this thesis, it was hypothesized that, depending on the season, the delivery of ‘old’ water to streamflow could be attributed to the transmissivity feedback mechanism and ‘old’ water movement through near-surface macropores. Studies with the explicit goal of identifying and quantifying specific subsurface flow mechanisms and the conditions that promote those mechanisms are therefore needed to verify those findings in other environments and hence better understand how ‘old’ water is rapidly delivered to streams.

5.3 References

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