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Field-scale water balance closure in seasonally frozen conditions

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7 Abstract

8 Hydrological water balance closure is a simple concept, yet in practice it is usually impossible to
9 measure every significant term independently in the field. Here we explore field scale water
10 balance closure in a prairie pasture field site in Saskatchewan, Canada. The area is cold, flat and
11 semi-arid, with snowmelt-dominated runoff. Arrays of snow and soil moisture measurements were
12 combined with a precipitation gauge and flux tower evapotranspiration estimates. We consider
13 three hydrologically distinct periods: the snow accumulation period over the winter, the snowmelt
14 period in spring, and the summer growing season. Over two years studied (1st November 2012 to
15 31st October 2014), we saw similar snowpacks develop each winter result in markedly different
16 runoff responses during melt. This was attributed to different soil moisture conditions prior to the
17 snow accumulation period in each year. In the more typical year (2013), the snow pack mostly
18 infiltrates into the soil, and the water balance is dominated by vertical land-atmosphere exchanges.
19 However, in the wetter year (2013-2014), the snowpack was not absorbed as soil moisture, and
20 significant losses (i.e. deep or lateral fluxes) occurred in response to rainfall in the early growing
21 season. As a result, we were unable to close the water balance. In particular, we were unable to
22 quantify how the excess melt water was partitioned between lateral runoff and vertical soil
23 drainage leading to groundwater recharge. Shallow piezometers suggest groundwater recharge was



1 significant in the wet year, and was depression focused. It is concluded that models which use
2 physically-based process representations to partition the melt cannot be rigorously validated using
3 conventional field-scale measurements based on water balance residuals. Rather, models should
4 be constrained using direct observations, accounting for uncertainty, and there is a need to establish
5 which observations (what, where and when) are most effective at constraining the uncertainties in
6 the water balance components.

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8 **Keywords:** water balance closure; field scale; seasonally frozen soils; hydrological processes

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1 1 INTRODUCTION

2 Water balance closure has been described as the holy grail of scientific hydrology (Beven, 2006).
3 Beven suggests that the most important problem in hydrology in the 21st Century is providing the
4 techniques to measure integrated fluxes and storages at useful scales. In the current paper, we
5 define the problem of water balance closure as that of independently quantifying each term in the
6 water balance equation, such that the changes in storage within a specified domain and over some
7 time interval are adequately balanced by the net fluxes into/out of that domain over the same time
8 interval. As simple as this concept is, it has proven to be extremely hard to achieve in field studies.
9 For example, Mazur et al. (2011) reported a water balance closure study for a well-characterized,
10 intensively monitored artificial catchment, and were unable to close the water balance due to their
11 inability to quantify evapotranspiration and changes in storage. Natural heterogeneity of both water
12 fluxes and moisture states, which can vary at spatial and temporal scales that are beyond (or
13 beneath) our measurement capabilities, can make the task of observing complete water balance
14 closure seem like an enigmatic pursuit.

15 In this paper we present a case study from a heterogeneous pasture site in the Canadian prairies,
16 where we have quantified the various components of the water balance at the field scale, and
17 critically examine some of the simplifying assumptions which are often invoked when applying
18 water budget approaches in applied hydrology. The Canadian Prairie region lies in the southern
19 part of the provinces of Alberta, Saskatchewan and Manitoba, and makes up the northern portion
20 of the Great Plains region of North America. The hydrology of this region is markedly influenced
21 by the regional climate and geology, and at first glance appears to have a relatively simple water
22 balance. Much of the rainfall occurs during the growing season and is consumed by
23 evapotranspiration, resulting in very little surface runoff. Extensive past glaciations have blanketed



1 the region with a thick compacted till which has very low permeability (Keller et al., 1989),
2 resulting in relatively small interactions between the surface water and the underlying groundwater
3 regime (van der Kamp and Hayashi, 2009). As such, the water balance in this region is
4 conceptualized to be dominated by vertical exchanges of precipitation and evapotranspiration
5 between the soil and the atmosphere.

6 However, certain characteristics of the prairies region also make the hydrology exceedingly
7 complex and are likely to confound simple 1-d assumptions regarding the water budget. In
8 particular, the region is seasonally frozen, with long winters (4-6 months), featuring many
9 cryosphere-dominated hydrological processes. Approximately one third of annual precipitation is
10 snowfall, which is subject to extensive wind redistribution throughout the landscape (Pomeroy et
11 al., 1993; Pomeroy and Li, 2000) resulting in a spatially variable water input. During the spring
12 melt, spatially variable surface albedos and heat advection from snow-free to snow-covered areas
13 can cause differential rates of snow melt (Shook et al., 1993; Liston, 1995). Moreover, infiltration
14 into frozen soil has complex dependencies upon the antecedent moisture, the rate of melt, and local
15 topography (Gray et al., 2001), resulting in a highly variable spatial infiltration pattern (Hayashi
16 et al. 2003, Lundberg et al., 2016). Due to these factors, the annual snowmelt event typically
17 produces 80% or more of the annual local surface runoff (Gray and Landine, 1988). The
18 hydrological complexity of the landscape is also largely influenced by glacial and post-glacial
19 geomorphological processes which have imparted a tremendous degree of heterogeneity. Morainal
20 deposits, comprised of a variable mix of soil textures, are often topographically indeterminate and
21 consist of areas which are internally drained and infrequently contribute to stream flow (ZebARTH
22 and de Jong, 1989; Shaw et al. 2012). Interactions between the glaciers and underlying ancient
23 marine sediments have resulted in the glacial tills being rich in soluble salts (Henry et al., 1985;



1 Keller and van der Kamp, 1988), which in some cases have been concentrated near the surface,
2 creating both extensive and local soil salinization problems (Miller et al., 1981; Bedard-Haughn,
3 2009).

4 As a result of the aforementioned complexities, the prairie water balance is subject to spatial
5 variability of precipitation inputs, water storage characteristics, and evaporation patterns. Because
6 these hydrological characteristics do not necessarily vary at the same scales, the challenge of
7 properly quantifying the water balance terms becomes can be quite daunting. While observations
8 of all of the hydrological fluxes and states at large, i.e. useful (Beven, 2006), scales is desirable,
9 current measurement approaches do not yet fully permit this. The evaporative flux can be measured
10 over reasonably large scales (on the order of hundreds of meters) using the eddy covariance
11 technique, whereas the soil moisture status and drainage fluxes can generally only be measured at
12 point scales. Recent advances in remotely sensed soil moisture, such as the ground-based cosmic
13 ray neutron probe (Zreda et al., 2008) or satellite-based sensors such as those used by the Soil
14 Moisture and Ocean Salinity (SMOS) mission (Kerr et al., 2010) or the Soil Moisture Active
15 Passive (SMAP) mission (Entekhabi et al., 2010), can retrieve soil moisture estimates over
16 hundreds of meters to tens of kilometers. However, these observations are limited to the near-
17 surface, and need to be depth-scaled to the root-zone to be suitable for water balance studies
18 (Peterson et al., 2016). Adequately capturing field-scale variability using point scale measurement
19 techniques is an ambitious endeavor, requiring a very large number of samples (e.g. Grayson and
20 Western, 1998; Famiglietti et al., 2008; Brocca et al., 2010).

21 In this paper, we use a well-instrumented field site to quantify the magnitude of the water
22 balance components as they vary across three distinct seasons in the prairies, and evaluate whether
23 simplifying assumptions can be justified. **In Section 2 we provide a description of the study site**



1 and the instrumentation used. In Section 3 we present and discuss the results, focusing on the
2 measured components of the field scale water balance and their uncertainty. Finally Section 4
3 summarizes the insights and findings from this study.

4

5 2 METHODS

6 2.1 Field-scale water budget

7 We consider field scale to represent an area of the order of 500 m x 500 m, from the ground surface
8 to a depth of 1.6 m. This was the depth range that we were able to install neutron probes to monitor,
9 and is deep enough to capture all of the significant soil moisture dynamics at our site (see for
10 example Fig. 5 and 8). At this scale, storage terms include surface storage, ΔS_s , which includes
11 snow and ponded water, and subsurface (vadose zone) storage, ΔS_v , which is liquid and solid (ice)
12 soil moisture integrated over the root zone (taken to be 1.6 m). The field-scale vertical water
13 balance can hence be expressed as

$$\Delta S_s + \Delta S_v = P - ET - R - G - D \quad (1)$$

14 where all terms are in units of mm, and P is precipitation (solid and liquid phases), ET is
15 evaporation (including soil and free water evaporation, plant transpiration, and ice sublimation), R
16 is surface runoff leaving the field domain, G is drifting snow leaving the field domain laterally,
17 and D is vertical soil drainage at 1.6 m depth. The water table is located 3-5 m below ground
18 surface depending on location (the water table is shallower in topographic depressions) and time
19 of year (the water table is shallowest in the earlier summer after the melt period). Water table
20 dynamics are modest, but there will be lateral saturated flow processes occurring. Since the
21 saturated zone is well below the domain of our water balance, here we only consider vertical



1 drainage from the base of our soil layer. Lateral unsaturated subsurface flow may occur at local
2 scales due to changes in soil properties, but we do not expect these fluxes to be significant at field
3 scale. Hence, lateral subsurface fluxes are neglected. However, we will also discuss the infiltration
4 flux in this paper, since this is of great practical significance, even though it is an internal flux
5 within our domain.

6 In seasonally-frozen environments where winters are long and cold, processes in the summer and
7 winter are markedly different. A water year in this region is typically defined as from November
8 to October, such that snow accumulation and melt occur within the same water year. Annual water
9 balances are useful, but do not elucidate the important seasonal processes – in particular the storage
10 dynamics. **For a more rigorous analysis**, here we examine the water balance over three distinct
11 seasons:

- 12 • *The snow accumulation period*: starts from the first killing frost (<2°C), and ends at the
13 beginning of snowmelt;
- 14 • *The melt period*: in a typical year this starts from the peak snowpack, and ends when the
15 ground is completely snow free, typically 2-4 weeks sometime in March, April or May;
- 16 • *The growing season*: starts from the end of the melt period, and ends with the onset of the
17 snow accumulation period, roughly May to October.

18 In each period the nature of the individual components of the water budget is different. For
19 example, in the snow accumulation period, surface storage occurs as snow; in the growing season,
20 if it exists at all, it is as ponded water in ephemeral ponds which tend to dry out in early summer;
21 and in the melt period, it is a transition between these two. Snow drift, runoff and evaporation are
22 typically only significant in the snow accumulation season, the melt period, and the growing
23 season, respectively. This will be discussed in detail in Section 4.



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2 **2.2 Description of study site**

3 The instrumented field site ($51^{\circ} 22' 54''$ N, $106^{\circ} 24' 57''$ W) lies within a gauged sub-basin (Fig.
4 1) of the Brightwater Creek watershed, which is itself a sub-basin of the South Saskatchewan River
5 Basin. The gross area of the sub-basin defined by the Water Survey Canada gauge (05HG002) is
6 900 km², while the effective basin area, or that which would be expected to contribute flow to the
7 main stream channel during a flood with a return period of 2 years (Martin 2001), is just 282 km²
8 (Fig. 1a). Mean annual precipitation is about 330 mm (2009-2014), of which about 70 mm typically
9 falls as snow. Mean annual streamflow in the Brightwater Creek watershed is 4.95×10^6 m³ (1983-
10 2013), equivalent to 5.5 mm over the gross drainage area or 17.6 mm over the effective drainage
11 area. Annual runoff is therefore small by either measure. Streamflow is intermittent, and in most
12 years only occurs following snowmelt. The mean temperature in Jan and Jul is -12.9°C and 18.8°C,
13 respectively. The regional landscape consists of gently sloping glacio-lacustrine plains surrounded
14 by moraine deposits that have a rolling knob- and kettle-type topography (Miller et al., 1985). The
15 soils in the region are mainly Solonetzic and Chernozemic, and are mapped as Bradwell and
16 Asquith Associations as described by Ellis et al. (1968).

17 This local study area (500 m × 500 m) is located within a ~700 ha grazing pasture, which is
18 surrounded by fields cultivated in annual crops. Within the instrumented region, the topography
19 is undulating with a range in elevation of approximately 5 m. Vegetation consists of various
20 wheatgrasses (*Agropyron* sp.) and needle grasses (*Stipa* sp.) with patches of western snowberry
21 (*Symporicarpos occidentalis*), commonly referred to as buckbrush. Brush and grass communities
22 are interspersed in a spatial pattern on the order of 10's of meters. The texture of the soils within
23 the study area ranges from loam to clay loam.



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2 **2.3 Instrumentation**

3 A variety of measurements were used to characterize the field scale water balance from November
4 1, 2012, reported here until October 31, 2014. The evaporation flux, (mm), was obtained using the
5 eddy covariance technique. This consisted of a Campbell Scientific CSAT3 sonic anemometer and
6 a Campbell Scientific KH20 krypton hygrometer mounted on a scaffold tower located in the center
7 of the study area. The instruments were mounted at a height of 4.85 m above ground, and had a
8 representative measurement fetch of approximately 500 m (Burba, 2013). Raw data were collected
9 at a rate of 10 Hz, and latent heat fluxes ($Q_E = \lambda E$, where λ is the latent heat of vaporization or
10 sublimation) and sensible heat flux (Q_H) fluxes were calculated using Licor EddyPro software
11 (www.licor.com/eddypyro). Gaps in the flux data were filled using the Kalman filter and a dynamic
12 linear regression for recursive parameter estimation developed by Young and coworkers (Young,
13 1999; Young and Pedregal, 1999; Young et al., 2004), based on the relationship between latent
14 heat flux, available energy, and vapour pressure deficit. This gap-filling approach was evaluated
15 and recommended by Alavi et al. (2006) for filling gaps in latent heat flux data.

16 The available energy, consisting of the net radiation flux (Q_{NR}) and the ground heat flux (Q_G)
17 was measured at two locations within the eddy-covariance measurement footprint: representing
18 grass and brush surfaces. At the scaffold tower (grass surface), net radiation fluxes were measured
19 using a Kipp & Zonen CNR1 4-component radiometer; whereas, ~~at an auxiliary tripod (brush cover)~~
20 ~~located approximately 100 m from the scaffold,~~ net radiation fluxes were measured with a
21 Hukseflux NR01 4-component radiometer. At both locations 2 heat flux plates (Radiation Energy
22 Balance Systems model HFT3) were installed at a depth of 8 cm, and were laterally separated by



1 ~1 m. In order to calculate energy storage in the soil layer above the heat flux plates (ΔS_G), a single
2 volumetric water content sensor (Campbell Scientific CS650 dielectric permittivity sensor) was
3 installed at 5 cm depth, and a pair of averaging thermocouples were installed at 4 cm depth. The
4 available energy, calculated as the net radiation flux minus (plus) the amount of energy transferred
5 into (from) the soil, was averaged between the 2 locations.

6 The energy balance closure ratio (EBR),

$$EBR = \frac{\sum(Q_E + Q_H)}{\sum(Q_{RN} - Q_G - \Delta S_G)}, \quad (2)$$

7 was evaluated for each day of the growing season, which gave an average closure fraction of 0.72
8 and 0.74 in 2013 and 2014, respectively. These biases were corrected by forcing energy balance
9 closure using the measured Bowen ratio (*cf.* Twine et al., 2000) on a daily basis, which increased
10 the measured seasonal evaporation fluxes by 39% and 35% in 2013 and 2014, respectively. Biases
11 for the other seasons were not corrected since the evaporation fluxes over the frozen ground surface
12 were very small, and the turbulent heat fluxes are much more uncertain over snow (Helgason and
13 Pomeroy, 2012).

14 Precipitation, (mm), was measured by a Geonor T200-B weighing gauge. Biases in solid
15 precipitation (i.e. snow) measurements were corrected for undercatch using a wind speed-catch
16 efficiency relationship (Smith, 2007), and for liquid precipitation (i.e. rain) we assume a catch
17 efficiency of 95% for all rainfall measurements (Devine and Mekis, 2008). Precipitation bias-
18 correction leads an increase in measured precipitation of 19% and 13% in 2013 and 2014,
19 respectively.



Root zone soil water content and snowpack depth and density were measured at point locations in a crosshair pattern, comprising two perpendicular transects, centered on the flux tower (shown in the upper right corner of Fig. 1a). Water content was measured by a down-hole neutron moisture meter, model CPN 503DR Hydroprobe (CPN International Inc., Concord, CA). The blue pins are the neutron probe reading locations installed in June, 2012, and the yellow pins show new locations added in summer 2013. Volumetric soil moisture content (liquid water + ice) was measured at depth intervals of 0.2 m, from 0.2 m to 1.6 m below ground. Due to the problem of surface loss of neutrons, no readings shallower than 0.2 m were taken, meaning we may underestimate the changes in water content at the top of the soil profile. The change in soil water storage, ΔS_v , was calculated as the difference between any two moisture surveys, which were conducted with a time interval of around 2 weeks in the unfrozen period, and 2-3 times during the frozen period. Snowpack distribution along the long transect was investigated with a series of snow surveys during the snow covered period in late winter/spring of 2013 and 2014. Snow depth was measured at a distance interval of about every 2.0~3.0 m, and snow samples were taken from the neutron probe locations, i.e. every 50 m, using a core sampler (ESC30, Environment Canada, Canada) to determine snow density and calculate snow water equivalent (SWE). The change in snow water storage, ΔS_s , was calculated as the difference in the mean SWE between any two sampling dates. The topography of the long transect from northwest to southeast is shown in Fig. 1b.

Water table depths were monitored using piezometers, screened at a depth of around 5.5 m below ground, with level loggers (Solinst, Model 3001) at three locations along the northwest-southeast transect (#1, #2 and #3 in Fig. 1b). The one closest to the flux tower started collecting data in July 17, 2012, and the other two started in October 7, 2013. The measured water table depth



1 was corrected for changes in barometric pressure, measured at the flux tower, using the graphical
2 method for estimation of barometric efficiency proposed by Gonthier (2007).

3 Soil temperature was measured using Stevens Hydro-probes at three profiles, co-located with
4 the piezometers. At profile #1 (Fig. 1b) five probes were installed, at depths of 0.05 m, 0.2 m, 0.5
5 m, 1.0 m and 1.5 m below ground, and at the other two profiles seven probes were installed, at
6 depths of 0.05 m, 0.2 m, 0.5 m, 0.75 m, 1.0 m, 1.3 m and 1.6 m below ground. Measurements were
7 recorded every 30 minutes. The depth of the freezing front as a function of time was calculated by
8 interpolating the 0°C line from the soil temperature measurements.

9

10 **3 RESULTS AND DISCUSSION**

11 The hydrological conditions over the water years 2013 (November 1, 2012 to October 31, 2013)
12 and 2014 (November 1, 2013 to October 31, 2014), are shown in Fig. 2, and the quantified water
13 balance components for each season are presented in Table 1. Total annual precipitation was 292
14 mm (2013) and 385 mm (2014). In both years, snow accumulation started around the beginning of
15 November and snowmelt was complete by the end of April. The snowfall measured by the
16 precipitation gauge and expressed as SWE, was very similar each year: 75 mm (2013) and 72 mm
17 (2014) (Fig. 2b). Rainfall was 217 mm (2013) and 313 mm (2014). The sum of evaporation,
18 transpiration, and sublimation was 286 mm (2013) and 365 mm (2014). Both years had low
19 measured sublimation: 20 mm (2013) and 10 mm (2014). Most evapotranspiration occurred in
20 June, July and August. Water year 2013 was typical for the region in that soil moisture was
21 recharged following snowmelt, and then experienced drying over the summer months as ET
22 exceeded precipitation. Water year 2014 experienced a wet May-June, so that soil moisture



1 decreases were delayed to the latter part of the summer. In the following sections, the dominant
2 hydrological processes and water balance closure for each of the three seasons are described.

3

4 **3.1 Snow accumulation period**

5 During the snow accumulation period, we were able to measure precipitation, sublimation, snow
6 accumulation and changes in soil moisture. ~~Above the ground, the peak SWE, measured by snow~~
7 ~~surveys~~, exceeded ~~the~~ measured precipitation by 36 mm (2013) and 28 mm (2014), and, more
8 importantly, exceeded precipitation minus sublimation by 56 mm (2013) and 38 mm (2014). This
9 was expected, and is due to the lateral exchange of blowing snow, which is a characteristic of open
10 prairie environments. Whether this results in a net influx or efflux to a particular site generally
11 depends on the relative height of the local and surrounding vegetation, which can trap snow
12 (Pomeroy et al., 1993). The vegetation (grass and shrubs) within the pasture were apparently more
13 effective at trapping blowing snow than adjacent cropped fields with shorter stubble (usually less
14 than 15 cm). Within the pasture the spatiotemporal pattern of snow depth, ~~shown in Fig. 4~~ was
15 highly variable, with significantly larger snow amounts trapped at the tower (location M0, where
16 a fence and other infrastructure trapped snow) and within the brush vegetation (e.g. location S3).
17 Although the slight fluctuations in topography might also cause landscape-scale snow
18 redistribution, this appeared to be much weaker than the effect of vegetation.

19 During the snow accumulation period the soil freezes progressively from the surface downwards.
20 The maximum freezing depths were 1.3 m (2013) and > 1.6 m (2014), as shown in Fig. 2d. Given
21 similar snowpacks (which insulate the ground) and ~~similar surface temperatures~~, it is likely that
22 the differences in freezing depths were due to differences in soil moisture (discussed below). In



1 2014, wetter soils resulted in greater thermal diffusivity, because the increase in thermal
2 conductivity (tending to increase thermal diffusivity) was more significant than the increase in
3 specific heat capacity (tending to reduce the thermal diffusivity).

4 Figure 5 shows the change in root zone water content over the winter (from before soil freeze
5 up, to just before the soil thawed), from all available neutron probe measurements. There were
6 increases in water content over the winter, with larger increases nearer to the surface. The standard
7 deviation of change in soil water content also increases significantly nearer the surface, which
8 implies that the wetting process is non-uniform across the field. The net increase in the mean
9 (across different profiles) depth-integrated root zone soil moisture (ΔSv) was 24 mm (2013) and 8
10 mm (2014) (Table 1). Under frozen conditions the water content of the root zone can potentially
11 increase due to infiltration of mid-winter snow melt events (uncommon, but not unheard of in this
12 environment), or by upward migration caused by freezing induced hydraulic gradients (Hoekstra,
13 1966; Gray and Granger, 1986). However, there were no observations of mid-winter melt events
14 in this period (i.e. the temperature did not significantly rise above zero), so we do not believe
15 significant infiltration occurred, and upward moisture redistribution is a more plausible
16 explanation. Note that the water table dropped through the winter (Fig. 2), which could be partly
17 due to upward migration (Gray and Granger, 1986, Butler et al., 1996, Iwata et al., 2010).

18

19 **3.2 Melt period**

20 The observed timing and magnitude of snowmelt runoff in 2013 and 2014 at Brightwater Creek
21 (measured at gauging station 05HG002) is compared in Fig. 6. The timing of peak discharge in
22 both years is consistent with the timing of the depletion of the snowpack by melting. However, the



1 magnitude of the peak discharge in 2014 is much bigger than that in 2013, despite the snowpack
2 depths being comparable (Table 1). Runoff from our field site may or may not have directly
3 contributed to this watershed scale runoff (see the contributing area in Fig. 1a), but the local
4 infiltration/runoff behavior can still explain the differences seen at the larger scale. Peak SWE and
5 melt period changes in root zone soil moisture, were measured at coincident points on the transect
6 (Fig. 1b), and the results are shown in Fig. 7 (note that the transect was extended in 2014). In this
7 figure we show the peak SWE and the additional rain that fell during the melt period as inputs
8 (positive), and the increases in soil moisture (shown as negative numbers) are shown separately
9 for the snowmelt period (ΔS_{V1}), and the subsequent snow-free soil thaw period (ΔS_{V2}), which takes
10 considerably longer to complete (Fig. 2). If all of the snowmelt infiltrated and did not drain from
11 the soil, SWE and ΔS_V would be equal. In 2013, this was almost the case – most of the snowpack
12 infiltrated into the soil during the snowmelt period, with negligible increases in soil moisture after
13 the snowpack had gone. However, in 2014 there were minimal increases in ΔS_{V1} during melt,
14 implying that there was very little infiltration during melt, and hence the snowpack mostly went
15 to runoff. In 2014 significant infiltration occurred later, due to rainfall on April 15, 2014. These
16 results are consistent with the observed differences in basin-scale runoff for 2013 and 2014. The
17 assumption of negligible soil drainage during this period (=0 in Table 1) is important to these
18 interpretations. This is probably reasonable since to produce drainage, infiltrating water would
19 have to bypass > 1 m of frozen soil, and in both years, groundwater recharge, inferred from water
20 table rise (Fig. 2), does not commence until after the snowmelt period. However, this issue is
21 worthy of further investigation ~~in future studies~~.

22 For the same transect, soil water content profiles for pre-melt, post-snowmelt, and post-thaw
23 conditions are shown in Fig. 8. These observations show the strikingly different antecedent soil



1 moisture conditions in these two years, with dry antecedent conditions in 2013 and wet antecedent
2 conditions in 2014. It is well understood (Gray and Landine, 1988) that the infiltration capacity of
3 frozen soils depends strongly on the antecedent soil moisture. Saturated soils that freeze will tend
4 to produce impermeable conditions where infiltration is restricted and most of the snowmelt is
5 transformed into runoff, as happened here in 2014. Dry soils (or more specifically, soils where the
6 largest significant pores, which may be macropores, remain air filled) can maintain a high
7 infiltration capacity when frozen and the snowmelt infiltration can be significant, whilst runoff
8 may be negligible, as in 2013.

9

10 3.3 Growing period

11 Figure 9 shows **the observed water budget** for the growing period in 2013 and 2014. The two lines
12 represent the left (water storage change) and right (net precipitation, which is precipitation minus
13 evapotranspiration) hand sides of Equation (1), if runoff, soil drainage and snow drift are negligible.
14 In both years, we see that the cumulative ($P-E$) is well within the range of variability ~~in~~ the change
15 in ΔS_v , **but there are differences with the mean**. In 2013, soil moisture was supplied through snow
16 melt (see above) and was progressively depleted by evapotranspiration through the summer
17 months, with minimal rainfall inputs until a large event in late September (Fig. 2). **The discrepancy**
18 **in the water balance here implies that there is some input of water missing from the domain**. Whilst
19 it is conceivable that this water came from storage below the root zone (i.e. negative drainage, or
20 capillary flow) it is more likely due to errors in existing measurements, **i.e. underestimating,**
21 **overestimating, or overestimating the decline in associated with the simple averaging used to**
22 **upscale**. In 2014, due to significant rainfall in May and June, soil moisture continues ~~to~~ to increase
23 until July. Water balance errors are incurred from August to November, and are in the opposite



1 direction to the discrepancy seen in 2013 (i.e. the errors in late summer 2014 could be due to
2 overestimating, underestimating, or underestimating the decline). It is most likely that this error is
3 due to unaccounted for soil drainage, discussed below.

4 The water table is approximately 3-5 m below ground at this site. In 2013, the water table in
5 piezometer 1 (P1) rose steadily during the growing period, though the amount of rise was small, ~
6 20 cm (Fig. 2e). This is not consistent with our water balance based estimates of soil drainage
7 (there is no positive soil drainage in 2013). In 2014 the water table in P1 rose higher, ~ 50 cm rise
8 (Fig. 2e), implying that there was significantly more soil drainage this year, which is consistent
9 with our water balance. In 2014 we had two additional piezometers (P2 and P3) available along
10 a transect, including one piezometer (P3) located below a topographic depression. The response of
11 these three piezometers is shown in Fig. 10, along with the ground surface elevation. The water
12 table below the depression rose markedly more than in the upland piezometers, and peaked much
13 earlier (July). This is consistent with the depression focused recharge mechanism that has been
14 proposed for these environments (Hayashi et al., 2003). Blowing snow and surface water runoff
15 gathers in depressions and recharges the underlying groundwater, and this effect is enhanced when
16 the soils are frozen and impermeable. The elevated water tables below the depressions can then
17 lead to lateral redistribution in the saturated zone, beneath the uplands. Groundwater therefore
18 does not necessarily interact with the root zone beneath higher elevation land, with groundwater
19 recharge effectively going around it via the depressions and saturated zone. Moreover, since the
20 soil moisture in the depressions is likely very high – perhaps even saturated – changes in soil
21 moisture storage may be very small whilst the flux of water through these soils is large. This
22 phenomenon would not be captured by our soil moisture observations – e.g. profiles S5 and S6 in
23 2014 shown in Fig. 7e and highlights a limitation with using changes in soil moisture to infer



1 volumes of infiltration. This mechanism would have been more significant in 2014 due to
2 impermeable frozen soils and more surface runoff, but may also have been active in 2013, and
3 could explain the water table response.

4

5 **3.4 Annual water balances**

6 The measured water balance estimates in all periods include sizeable residuals (Table 1). Thus,
7 annual water balance estimates which are based upon residuals are subject to high levels of
8 uncertainty for two basic reasons: 1) cumulative measurement errors in water balance terms; and
9 2) assumptions about which water fluxes are present and how residual values are partitioned
10 between these fluxes may not be correct. For context, the residuals in this study (closure estimates
11 in Table 1), are seen to be of the same order of magnitude as the average annual basin runoff (~
12 18 mm). Of particular importance to the water balance estimates in this environment is the impact
13 of the bias corrections implemented to rectify the anticipated under-measurement of both
14 precipitation and evapotranspiration. In this case, empirical adjustments to the annual precipitation,
15 intended primarily to correct for snow falling in windy conditions, amounted to 47 mm in 2013
16 and 44 mm in 2014. Similarly, adjustments to the latent heat flux, associated with forcing energy
17 balance closure, accumulated an additional 80 mm of evapotranspiration in 2013 and 95 mm in
18 2014. Fortunately, these large magnitude flux adjustments, which contain considerable uncertainty,
19 act in opposite directions and partially cancel each other out. From an accounting perspective, the
20 ability to close the water balance based on measurements alone is far from satisfactory. However,
21 the situation does highlight a defining challenge of semi-arid environment water balances, where
22 small differences between precipitation and evaporation are pivotal, and **accumulated errors limit**
23 **the accuracy of which other fluxes (i.e. runoff, drainage, infiltration) can be estimated.**



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2 **4. SUMMARY AND CONCLUSIONS**

3 In this study we have used a suite of relatively standard, albeit unusually comprehensive,
4 instrumentation to explore the field scale, root zone water balance. Due to the local climate, the
5 year is split into three periods, each summarized in the following paragraphs.

6 During the winter, i.e. the snow accumulation period, we were unable to close the water balance
7 because we could not measure the fluxes of blowing snow or upward soil moisture redistribution,
8 both of which are significant. However, for practical purposes, the most important measurement
9 during this period is the peak SWE in the snowpack before melt, which can be measured by a snow
10 survey, and can capture the pre-melt spatial variability. This measurement negates the requirement
11 to capture blowing snow and sublimation, and eliminates uncertainties in the measured solid phase
12 precipitation. The measured SWE can be used to validate or drive hydrological models, depending
13 on the modelling objective. We also see that the increase in soil moisture over the winter is
14 significant, likely due to upward moisture redistribution, and so measuring soil moisture prior to
15 melt, if possible, is valuable.

16 The snowmelt period in the Canadian prairies, as illustrated by this field study, strongly
17 dominates the subsequent hydrological processes. A fundamental challenge is to predict how the
18 melting snowpack will be partitioned between runoff and infiltration, which is a strong determinant
19 of flood risk and soil moisture availability. Our observations are consistent with past studies that
20 have highlighted the importance of antecedent soil moisture in generating runoff. We also have
21 observations of groundwater that are consistent with the depression focused recharge mechanism.
22 In terms of water balance closure, we were unable to obtain independent field scale measurements



1 of runoff and drainage. The magnitude of the residual flux is uncertain, and the partitioning of this
2 flux between runoff and drainage is also uncertain. This should be accounted for when using water
3 balance residuals to constrain models.

4 From a water balance perspective, the growing season was the least problematic in this study.

5 Here, the important question for agricultural production is how much water is available for use by
6 plants. A simple vertical water balance (rainfall minus evaporation) seems to adequately explain
7 the changes in moisture. The residuals are -28 mm (2013) and $+28\text{ mm}$ (2014), which is about 10%
8 of precipitation. This is likely acceptable for tracking short term (e.g. one season) hydrological
9 processes. However, longer term groundwater recharge and solute transport processes (e.g. salts
10 and nutrients) are driven by fluxes that may be much smaller than these residuals, but are important
11 over 10s to 100s of years. In this case, we are limited by the accuracy of our precipitation and
12 evapotranspiration estimates, and should be very cautious with quantitative estimates that do not
13 consider uncertainties in these observations.

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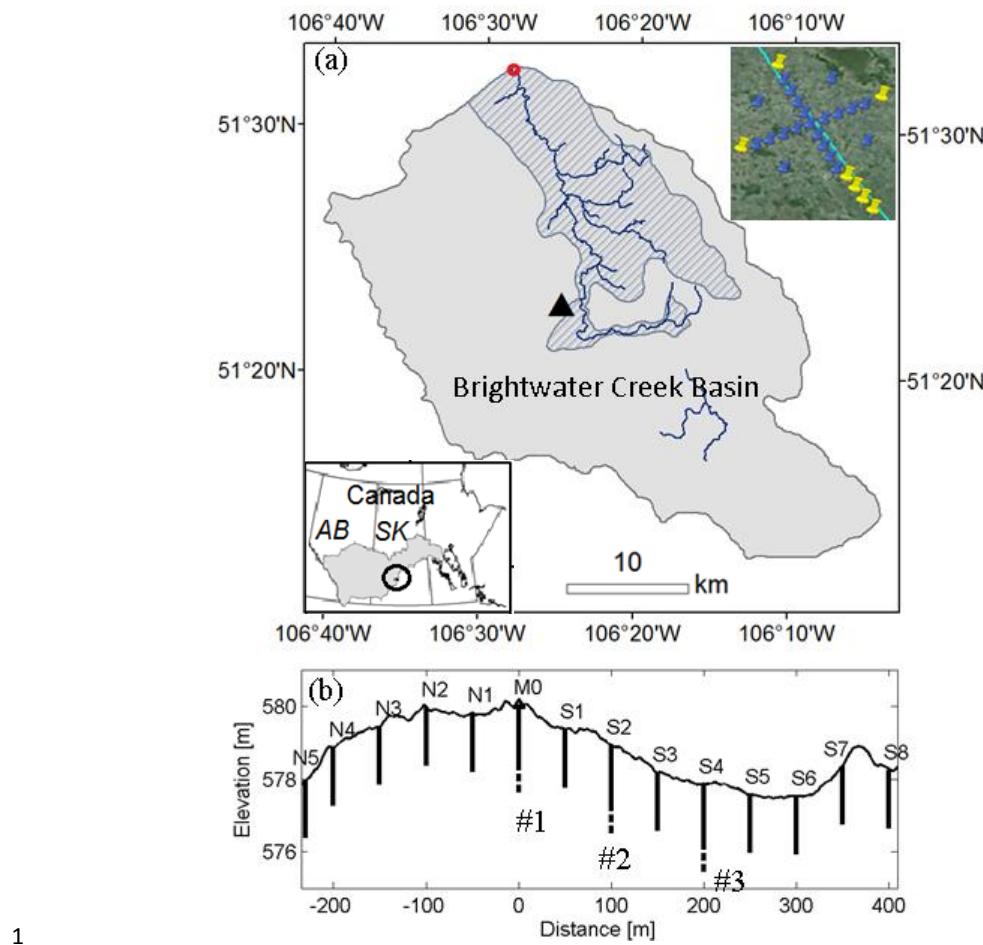


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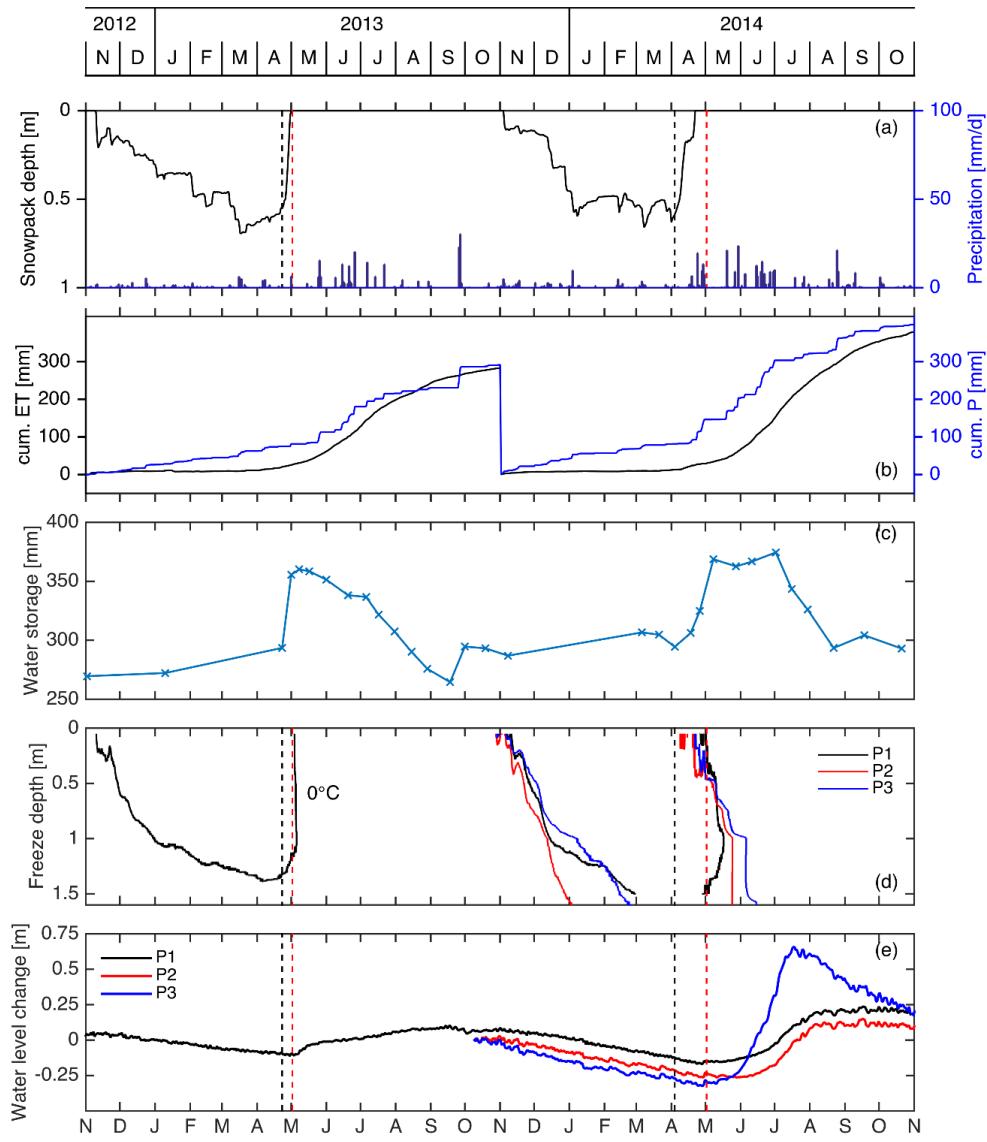
1 Table 1. Components of the water balance for each season in 2013 and 2014. The closure estimates
 2 are our best judgment of where the water balance residuals come from, but are uncertain. Res.
 3 represents water balance residuals that are mostly likely due to errors in other measurements.
 4 Evapotranspiration (ET), runoff (R), snow drift (G) and soil drainage (D) are all positive for losses
 5 from the control volume.

Seasons	Observations [mm]				Closure estimates [mm]			
	ΔS_s	ΔS_v	P	ET	R	G	D	Res.
Snow Acc. '13 (Nov. 1 - Apr. 21)	111	24	75	20	0	-56	-24	0
Melt '13 (Apr. 22 - May 6)	-111	67	7	11	40	0	0	0
Grow '13 (May 7 - Nov. 6)	0	-73	210	255	0	0	0	-28
Snow Acc. '14 (Nov. 7 - Apr. 2)	100	8	72	10	0	-38	-8	0
Melt '14 (Apr. 3 - Apr 24)	-100	19	33	15	99	0	0	0
Grow '14 (Apr 25 - Oct. 22)	0	-32	280	340	0	0	28	0

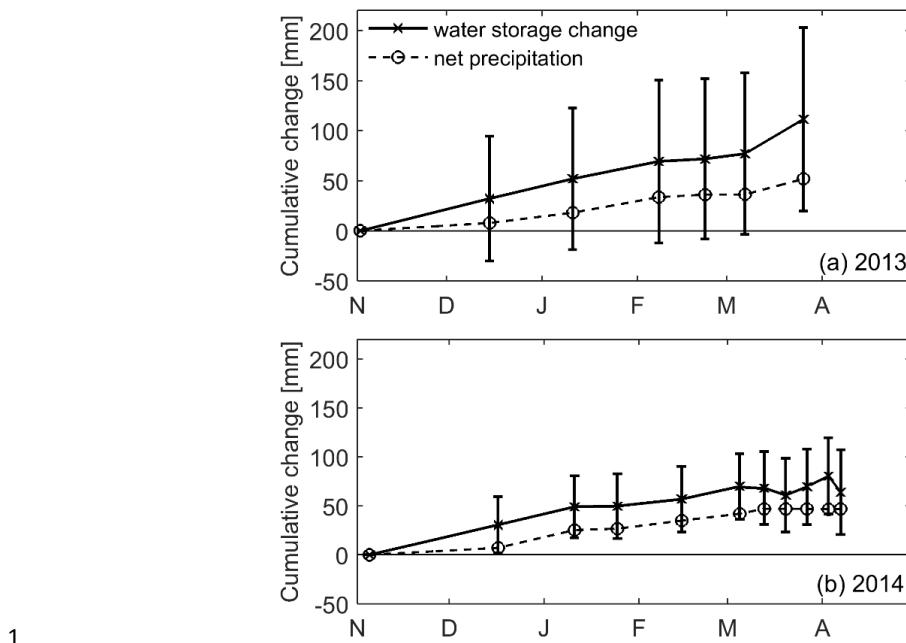


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2 Figure 1 Brightwater Creek sub-basin in Saskatchewan River basin (effective drainage area shown
3 by hatching) and locations of measurements. (a) Flux tower (triangle) and the schematic
4 distribution of neutron monitoring locations (right upper corner); Red circle: discharge
5 measurement location. (b) Instrumentation along the long transect in the (a): neutron probe access
6 tubes (N5, N4, N3, N2, N1, M0, S1, S2, S3, S4, S5, S6, S7, S8), **snow survey**, and three 6-m
7 piezometer boreholes (#1, #2 and #3).

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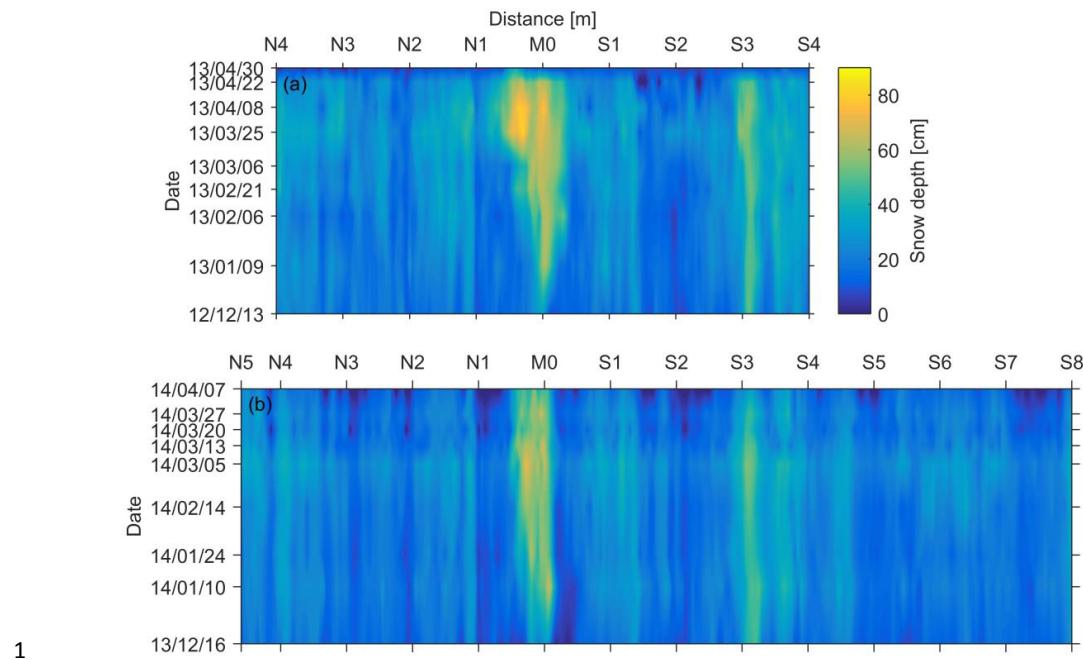
2 Figure 2 Fluctuation of the major variables in vadose zone hydrology during the years of 2013 and
3 (a) Precipitation and snowpack depth measured at the flux tower. (b) Yearly cumulative
4 change of evapotranspiration (ET) and precipitation (P). (c) Average soil water storage change in
5 shallow vadose zone (neutron probe data). (d) Surficial soil freezing above groundwater table at
6 three locations (P1, P2 and P3). (e) Seasonal fluctuation of groundwater level. Black and red
7 dashed lines are the start and end of the snowmelt periods.



1 Figure 3 **Non-closure** of the vertical field-scale water balance during the snow accumulation period
2 in hydrological years of 2013 (a) and 2014 (b). Note the error bar just indicates spatial variance
3 (same as Fig. 5 and Fig. 9).

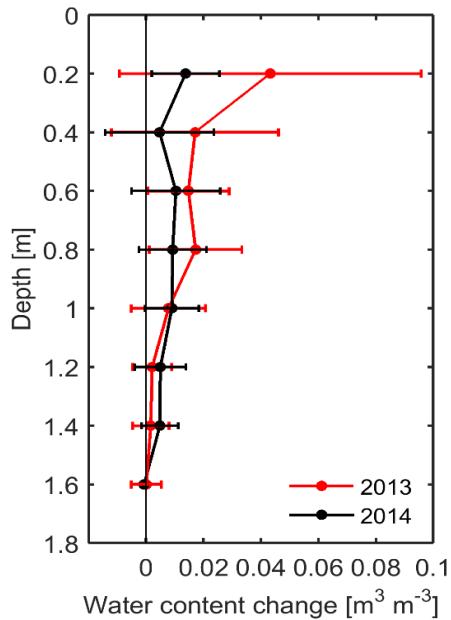
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2 Figure 4 Spatio-temporal variation of snowpack depth along the snow survey transect in 2013 (a)
3 to 2014 (b).

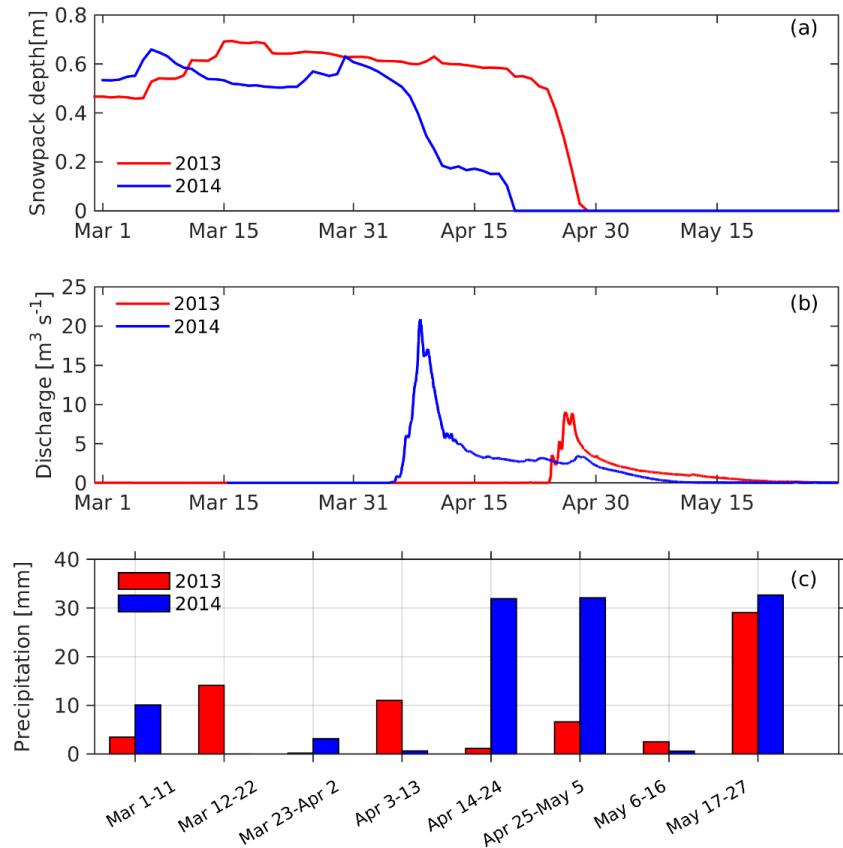
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2 Figure 5 Over-winter change in water content with depth below ground. Dots indicate the mean
3 over all neutron probes, and bars indicate standard deviation from the mean. In water year 2013,
4 the pre-freeze measurement was taken on November 1st, 2012, and the pre-snowmelt measurement
5 on April 22nd 2013. In water year 2014, the pre-freeze measurement was taken on November 7th,
6 2013, and the pre-snowmelt measurement on April 3rd, 2014.

7

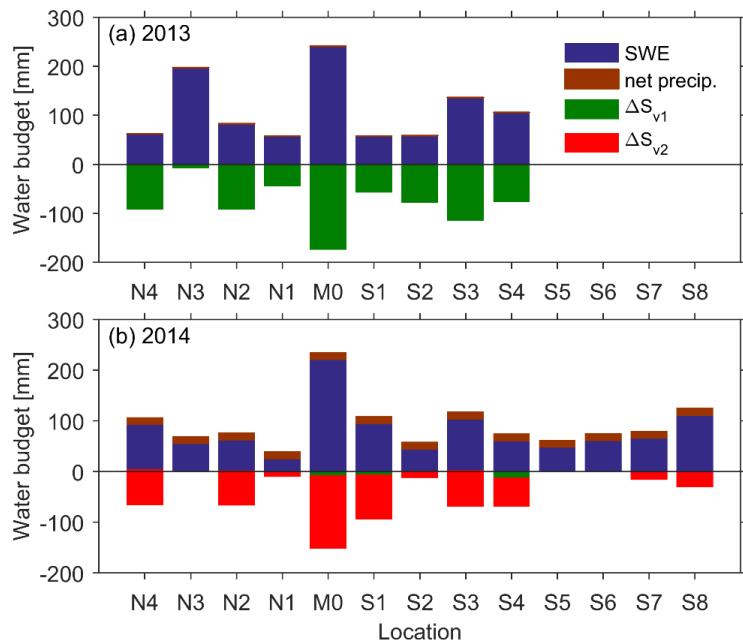


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2 Figure 6 Contrasting snowmelt processes observed in 2013 and 2014. (a) Snowpack depth. (b)
3 Hydrograph of the Brightwater Creek. (c) Snowfall and rainfall with 10-day interval during melt
4 period.



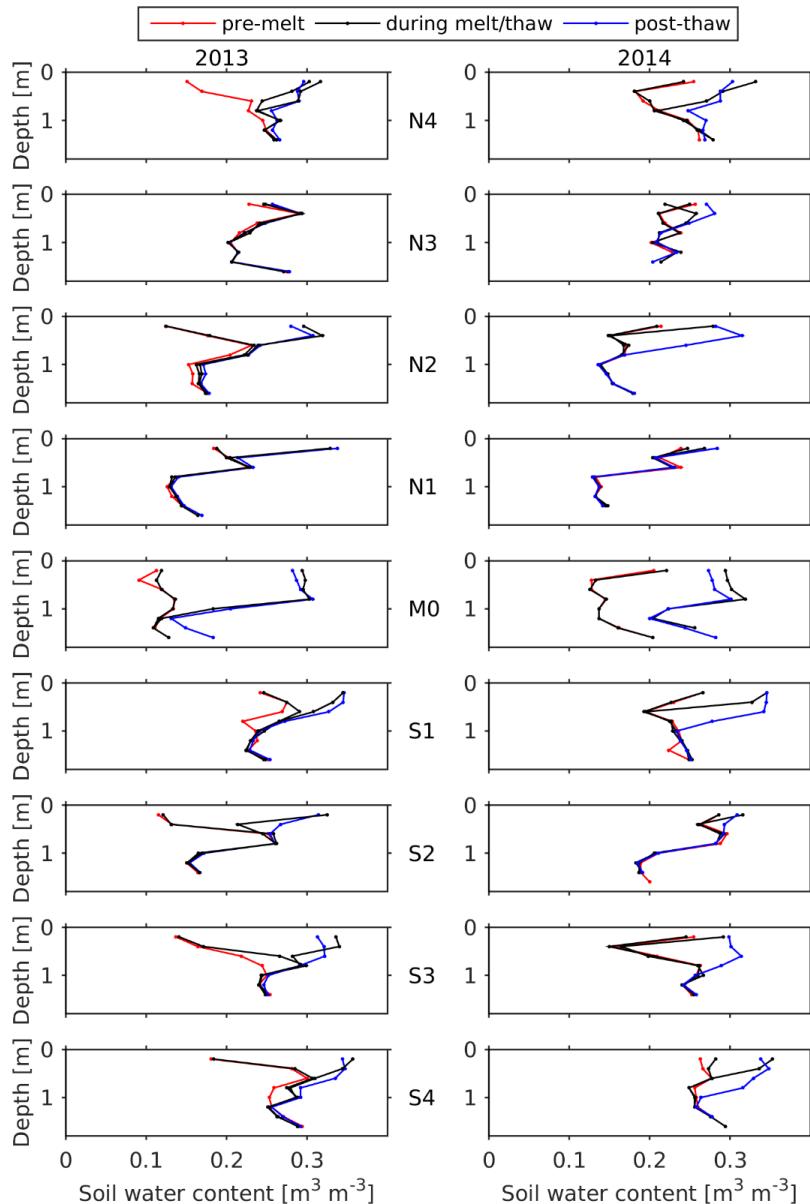
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3 Figure 7 Spatio-temporal variations of soil water storage change in the shallow vadose zone along
4 the neutron probe transect during the melt period (between black and red dashed lines in Fig. 2).
5 SWE: measured maximum snow storage; net rainfall: cumulative difference between precipitation
6 and evapotranspiration; ΔS_{v1} and ΔS_{v2} : soil water storage change during post-snowmelt and post-
7 thaw periods.

8

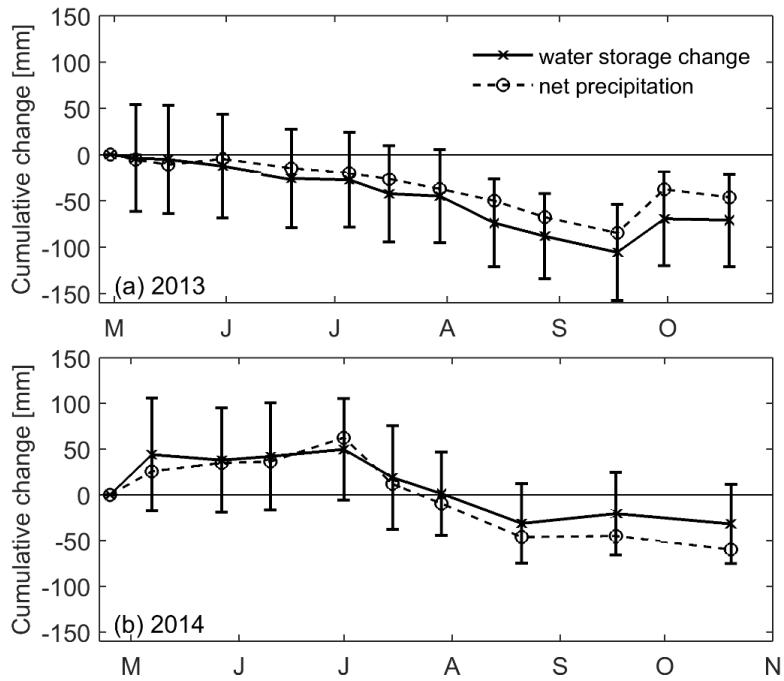


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2 Figure 8 Spatio-temporal variation of water content in shallow vadose zone at different locations
3 (Fig. 1b) along the Neutron Probe reading transect during the pre-melt (red), post-snowmelt
4 (black) and post-thaw (blue) in 2013 (left panel) and 2014 (right panel).



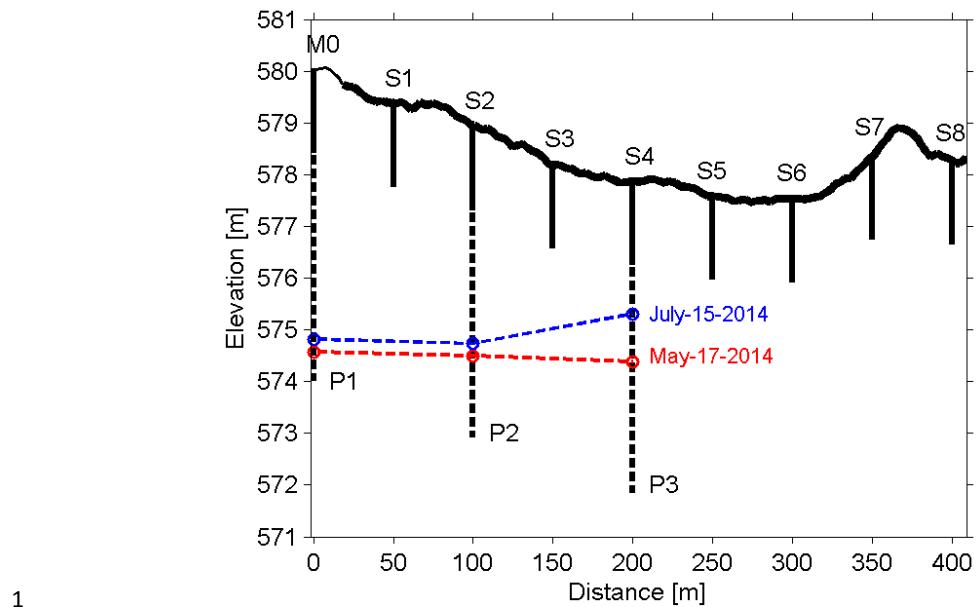
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3 Figure 9 Non-closure of the vertical field-scale water balance during the growing period in
4 hydrological years of 2013 (a) and 2014 (b).

5



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2 Figure 10 Groundwater rise along the slope during the early growing period in 2014.