Assessing hydrological sensitivity of grassland basins in the Canadian Prairies to climate using a basin classification–based virtual modelling approach

C. Spence[1*], Z. He[2], K.R. Shook[2], B.A. Mekonnen[3], J.W. Pomeroy[2], C.J. Whitfield[4], J.D. Wolfe[5]

* Corresponding author: Christopher Spence (chris.spence@canada.ca)

[1] Environment and Climate Change Canada, Saskatoon, Saskatchewan, Canada

[2] Centre for Hydrology, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

[3] Golder Associates, Calgary, Alberta, Canada

[4] School of Environment and Sustainability, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

[5] Saskatchewan Ministry of Environment, Regina, Saskatchewan, Canada

*corresponding author: chris.spence@canada.ca

Abstract

Significant challenges from changes in climate and land-use face sustainable water use in the Canadian Prairies ecozone. The region has experienced significant warming since the mid 20th Century, and continued warming of an additional 2°C by 2050 is expected. This paper aims to enhance understanding of climate controls on Prairie basin hydrology through numerical model experiments. It approaches this by developing a basin classification–based virtual modeling framework for a portion of the Prairie region, and applying the modelling framework to investigate the hydrological sensitivity of one Prairie basin class (High Elevation Grasslands) to changes in climate. High Elevation Grasslands dominate much of central and southern Alberta and parts of southwestern Saskatchewan with outliers in eastern Saskatchewan and western Manitoba. The experiments revealed that High Elevation Grasslands snowpacks are highly sensitive to changes in climate, but that this varies geographically. Spring maximum snow water equivalent in grasslands decreases 8% per °C of warming. Climate scenario simulations indicated a 2°C increase in temperature requires at least an increase of 20% in mean annual precipitation for there to be enough additional snowfall to compensate for enhanced melt losses. The sensitivity in runoff is less linear and varies substantially across the study domain; simulations using 6°C of warming and a 30% increase in mean annual precipitation yields simulated decreases in annual runoff of 40% in climates of the western Prairie but 55% increases in climates of eastern portions. These results can be used to identify those areas of the region that are most sensitive to climate change, and highlight focus areas for monitoring and adaptation. The results also demonstrate how a basin classification–based virtual modeling framework can be applied to evaluate regional scale impacts of climate change with relatively high spatial resolution, in a robust, effective and efficient manner.

Key words: Prairie, basin classification, virtual experiments, climate change, snow, runoff

1

Introduction

Hydrological models are essential tools to understand hydrological processes and function at the basin scale, and can also be used to diagnose how specific hydrological processes control catchment responses to change (Rasouli et al., 2014). Modelling a specific basin to evaluate processes or to simulate the effects of change entails large computational and labour costs and requires observations of the basin response with sufficient spatial and temporal coverage. Modelling of many individual basins is not efficient when attempting to predict regional responses to changes in climate and/or land-use. Basin classification can regionalize hydrological model outputs, based on the assumption that basins can be classified by their characteristics and that basins of the same class respond similarly to changes in climate inputs or their landscapes (e.g., McDonnell and Woods, 2004; Wagener et al., 2007). Parameterizing a model based upon a representative or stylized basin of a given class allows the output to be considered representative of all basins of that class. This assumption facilitates regionalization as it does not necessarily require simulating the distinctive characteristics of every basin, reducing cost and time required for large domain studies. Such a regionalization approach can be used to assess the sensitivity of large diverse areas to stressors, such as land-use and climate change.

One such region is the Canadian Prairie ecozone, that portion of the Great Plains of North America that includes southern parts of the provinces of Alberta, Saskatchewan, and Manitoba and Treaties 1, 2, 4, 6 and 7 in Western Canada (Spence et al., 2019). This region has a cold sub-humid to semi-arid climate and was covered by grassland and sparse woodlands until colonization of the region promoted conversion of large expanses of natural grassland to

cultivated agriculture in the late 19th and early 20th centuries. The region's geomorphology was

65   formed by glacial and post-glacial processes which left numerous internally drained depressions

and poorly defined drainage networks. Most of the Canadian Prairies is in the Saskatchewan-

Nelson River Basin, but relatively little runoff is provided to the major rivers that traverse the

region downstream of their mountain headwaters. Local streams and prairie-derived rivers often

have intermittent and highly variable streamflow. These streams are important local sources of

70   freshwater and are often managed to provide farm, agricultural and municipal water supply and

support natural lakes and reservoirs (Pomeroy et al., 2005). Because they connect to larger

systems only intermittently, a small headwater–basin scale approach is necessary to generate

information about how their behaviour might be impacted by the aforementioned stressors.

75   Western Canada, including the Canadian Prairies, has been subject to substantial climate

warming since the mid 20th century (Johnson et al., 2005; Bush and Lemmen, 2019). Prairie

precipitation trends indicate more rain and less snow in the spring and fall (Shook and Pomeroy,

2012) and runoff generation has been shown to be shifting from snowmelt- to rainfall-driven in

eastern Saskatchewan (Dumanski et al., 2015). Recent analysis of hydrometric stations across the

80   region identified sub-regional trends in streamflow associated with drying in the west and south

and wetting in the east and north, associated with physical landscape characteristics and climate

(Whitfield et al., 2020). However, it is difficult to attribute streamflow response solely to

climate change because of impoundment of streams, widespread changes in agricultural practices

and wetland drainage since the 1950s (Ehsanzadeh, 2016). Wetland drainage has become

85   widespread in portions of the region (van Meter and Basu, 2015) and the loss of depressional

storage capacity associated with drainage enhances streamflow volumes (Tiner, 2003; Wilson et

al., 2019) and may alter the frequency, timing, and duration of regional streamflow (Ehsanzadeh et al., 2012; Spence and Mengistu, 2019). Extrapolating intensive studies of wetland drainage impact in individual basins (Wilson et al., 2019) can be challenging, because basin response is a

90   function of wetland distributions that control contributing area dynamics (Stichling and Blackwell, 1957; Shaw et al., 2012; Shook and Pomeroy, 2011; Haque et al. 2017, Spence and Mengistu, 2019). It is uncertain how hydrological fluxes and states in Canadian Prairie basins will respond to continued climate change and wetland drainage. The statistical modelling and small basin modelling studies cited here have provided an excellent foundation, but an improved

95   approach is needed to evaluate how changes in climate and agricultural practices impact hydrological regimes more broadly across the region.

Here, a classification-based virtual modeling framework is proposed as a means to examine hydrological sensitivity to different climate, land-use and wetland drainage. In this approach,

100 each basin class is modelled in a virtual manner (Weiler and McDonnell, 2004; Armstrong et al., 2015); as a synthetic or generic basin with characteristics defined by the average or typical condition of all basins from the same class. In this way, the basin characteristics can be manipulated to determine how a typical basin may respond to change. There is evidence that such an approach is viable, as virtual experiments have been used to evaluate hydrological

105 response to different conditions (Di Giammarco et al., 1996; Horn et al., 2005; Dunn et al., 2007; Mallard et al., 2014; Seo and Schmidt, 2013, Lopez-Moreno et al., 2020), identify factors influencing hydrological processes (e.g., Weiler and McDonnell, 2004), and study hydrological controls on water chemistry (Weiler and McDonnell, 2006). This paper aims to demonstrate the utility of a basin classification–based virtual modelling approach for assessing the sensitivity of

4

110   Canadian Prairie catchments to climate. Two steps were taken to achieve this objective: (1)

development of a robust class-based virtual basin model for a portion of the Canadian Prairie

and; (2) exploration of virtual basin sensitivity of hydrological response to climate within this

portion of the Canadian Prairie. This work provides a foundation to extend the virtual basin

modelling approach more broadly across the Canadian Prairie to assess response to climate and

115   land management scenarios.

Methodology

Framework of classification-based virtual basin modeling

A basin classification–based virtual modelling platform has three main components: (1) a

120   classification analysis to derive virtual basin characteristics; (2) parameterization and evaluation

of a hydrological model of the virtual basin and (3) application of the model to evaluate response

to multiple scenarios (Figure 1). This approach was selected because other regionalization

approaches had particular disadvantages. For example, the low density of gauged non-regulated

streams in the region make extrapolation of observed streamflow data highly uncertain. An

125   alternative was to use gridded climate and climate scenario data to model numerous individual

catchments across the region and then regionalize those results. However, we were unable to

find a gridded climate product that performed as well over a multi-decadal period as the climate

station data that was ultimately selected.

130   *Basin Classification*

The classification of Canadian Prairie basins was based on the analyses of Wolfe et al. (2019),

which identified over 4000 basins, each approximately $100 \ \text{km}^2$ in area, as being in one of seven

5

broad classes, based on a suite of physio-geographic characteristics (Figure 2). A more thorough

detailed approach is needed within this region because methods such as Knoben et al. (2018) are

135   too coarse to differentiate functional differences over this region. This classification was

intended to elucidate and define these differences for the purpose of hydrological model design

and parameterization. The basin delineations used in the study were taken from the HydroSHEDs

dataset (Lehner and Grill, 2013), which provides geographically contiguous delineations of

basins for the world including the Prairie ecozone. Physio-geographic characteristics, including

140   climate, streamflow, geology, topography, wetland distribution, and land cover, among others,

were compiled and used to classify basins that would be expected to respond in a hydrologically

coherent manner (Wolfe et al., 2019). Urban areas and large lakes were excluded from the

analysis. A revised classification that excluded climate parameters but retained the streamflow

regime parameters was conducted and is used herein, following the same Hierarchical

145   Classification of Principal Components (HCPC) approach. This was done because including

climate could bias the results of the climate scenarios by "hard wiring" some functionality into

the model parameters. The influence of recent climate, historical climate change and how this

interacts with Prairie biogeophysical features and wetland drainage to influence basin hydrology

behaviour has been recently explored by Whitfield et al. (2021).   They showed the coherence

150   between climate effects and biogeophysical and drainage effects on basin classification.

Therefore, bias could be introduced from a classification that includes climate when applying

future climates to force the virtual basin model. For example, a current climate classification

might create a virtual basin class that precluded semi-arid basins, this would become non-

functional with projected increases in precipitation and temperature for the region which could

155   change to be semi-arid. By excluding the direct impact of climate in the classification and

6

restricting it to topography, biogeophysical features and drainage systems the new climates of

Prairie basins can be more fully explored.

The High Elevation Grasslands (HEG) class (Figure 2) was selected for the development of the

160   virtual basin model. This class featured 751 basins with an average size of ~100 km^2. Median

basin characteristics, including land cover fractions, basin slope and elevation, and soil type,

were determined and used for virtual basin model parameterization. Wetland areas derived from

the Global Surface Water maximum water extent dataset ([https://global-surface-](https://global-surface-water.appspot.com/download)

[water.appspot.com/download](https://global-surface-water.appspot.com/download)) were used to determine the shape and scale parameters of a

165   generalized Pareto distribution (Shook et al., 2013, Table 1). These parameters were used to

characterize the wetland complex within the virtual model, where the wetland complex is

represented by multiple individual wetlands of different sizes.

Model set-up and parameterization

170   The Cold Regions Hydrological Modelling platform (CRHM) was selected to develop the virtual

basin model, as CRHM is particularly suited for simulating the hydrology of the Canadian

Prairies. CRHM is a modular, process-based, spatially semi-distributed hydrological model,

which includes the key cold regions and warm season hydrological processes that operate in

western Canada and elsewhere (Pomeroy et al., 2007). With the correct suite of modules, each

175   representing a key hydrological process, CRHM has proven very capable of representing prairie

hydrological processes and accurately emulating water fluxes in this landscape (Fang and

Pomeroy, 2009; Fang et al., 2010). Most of the process modules in CRHM, particularly the

surface processes, are strongly physically-based, and hence CRHM does not require model

calibration, making it suitable for simulations under nonstationary conditions. As a virtual basin

180   has no specific location, it cannot be calibrated to field observations from a particular basin, so

setting parameters based on regional hydrological studies rather than calibration is advantageous.

The virtual basin area was set to 100 km^2, which aligns with the average basin size used in the

classification. The virtual basin was divided into hydrological response units (HRUs), each of

185   which has a single set of parameter values and for which water budgets are calculated. The HRU

sequence by which runoff is routed follows a catena of land cover from cultivated fields at the

highest elevation, followed by grasslands, shrublands, woodlands, and wetlands at successively

lower elevations. Areas were set according to the median for that land cover observed across all

HEG basins. Summer-fallow fields, whilst also included, cover a very small area (<1%; Table 1).

190

The virtual basin was separated into non-wetland, and wetland catenas (Figure 3) according to

effective and non-effective fractions of HEG basins, respectively. The first, 'non-wetland' catena

portion (i.e., cultivated, grassland, shrubland, woodland HRUs) of the basin (~67% of area) is

considered to contribute flows directly to the HRU outlet (stream channel). Runoff from the

195   'wetland' catena portion of the virtual basin (~33% of area) features a wetland complex HRU

within a landscape catena following a sequence from cultivated, to grassland, shrubland, and

woodland HRUs (Figure 3). Runoff is routed through a wetland complex comprised of 46

individual wetlands; their areas follow a . The selection of 46 wetlands was based on the work of

Shook and Pomeroy (2011) and Shook et al. (2013) who found that this was the minimum

200   number of wetlands to represent crucial hydrological behaviour (e.g., hysteresis) that controls the

dynamic area contributing flow downstream for a Prairie basin of similar size. The total wetland

8

area was based on the fractional area from basin classification. The distribution of this area among the 46 wetlands followed a generalized Pareto distribution as it has been used successfully to characterize the distribution of open water areas in similar landscapes in North

205   Dakota and central Saskatchewan (Zhang et al., 2009; Shook et al., 2013).

The Prairie Hydrological Model (PHM), created using CRHM (Pomeroy et al., 2010, 2012), was used to simulate virtual basin hydrological response. PHM includes a specific set of physically-based modules linked sequentially to represent the dominant hydrological processes for the

210   virtual basin. The Observation module reads in meteorological data and calculates the precipitation phase using a maximum temperature threshold of 0°C for 100% snow and a minimum temperature threshold of +2°C for 100% rainfall and distributes these to each HRU along with lapse rates for temperature and precipitation and temporal interpolation to adjust observation time steps to hourly data. The Radiation module outputs incoming shortwave

215   radiation to slopes, longwave radiation, and net radiation. The Canopy module (Sicart et al., 2004; Ellis et al., 2010) was used to adjust the effects of vegetation canopies on sub-canopy net radiation reaching underlying snow. The leaf area index was assigned a value of 0.4 for woodland HRUs, which is a typical value for aspen trees during winter (Pomeroy et al., 1999). For other HRUs, a minimum value of 0.001 was set to simulate the canopy effects of Prairie

220   vegetation (crop residue, grass) on radiation for snowmelt. The values of the initial albedos for bare ground and fresh snow were set to 0.17 and 0.85, respectively (Table 1), based on suggested values by Armstrong et al. (2008) for summer conditions and Male and Gray (1981) for snow.

9

To simulate snow redistribution processes, the Prairie Blowing Snow Model (PBSM) (Pomeroy

225 and Li, 2000), and Walmsley's windflow model (Walmsley et al., 1989) modules were used.

Fetch distances were set using values recommended in Pomeroy et al. (2007). Fetch distance was

set to 1000 m for exposed sites such as cultivated and fallow HRUs. A 500 m fetch was assigned

for grassland HRUs. For other HRUs, a 300 m fetch was selected. The distribution factor

parameterizes the allocation of blowing snow transport from aerodynamically smoother (or

230 windier) HRU to aerodynamically rougher (or calmer) ones and was selected according to the

prairie landscape aerodynamic sequencing of Fang and Pomeroy (2009). Vegetation height, stalk

density, and stalk diameter were set to represent the values for the prairie environment during fall

and winter (Table 1) (Pomeroy and Li, 2000).

235 The Energy-Budget Snowmelt Model (EBSM) (Gray and Landine, 1988) module includes

algorithms applicable to the Canadian Prairies and was used to simulate snowmelt by calculating

the balance of radiation, sensible heat, latent heat, ground heat, advection from rainfall, and

change in internal energy. Infiltration to unfrozen and frozen soils was calculated by the Prairie

Infiltration module with algorithms based on Ayers (1959) and Gray et al. (1985), respectively.

240 Actual evapotranspiration was simulated using the Penman-Monteith equation (Monteith, 1965),

and evaporation from typically saturated surfaces subject to advection, such as wetlands and

stream channels, was calculated using the Priestley and Taylor equation (Priestley and Taylor,

1972).

245 The Soil module calculates the water balance for the soil column, which is divided into two

layers: the top layer (called the recharge zone) and the lower layer. The depth of the recharge

layer was set at 1.4 m as this is typically a zone of higher hydraulic conductivity in the glacial till derived soils of the region (Brannen et al., 2015). Wolfe et al. (2019) identified the predominant HEG soil texture as loam, which was then assigned as the soil type for all HRUs. The

250   Muskingum routing method was used for the routing amongst HRUs. The routing length for each HRU was calculated using the modified Hack's Law length-area relationship, which was derived from a previous CRHM-PHM modeling study of Smith Creek in Saskatchewan (Fang et al., 2010; Pomeroy et al., 2010).

255   *Model application*

To ensure that the role of climate variability across HEG was captured in streamflow simulations the model was run over a 46-year baseline period (1960–2006) driven using data from seven locations. The locations were within and nearby the geographical extent of the HEG classification, and represented the variation in climate across the region (Figure 2; Table 2). The

260   mean annual average temperature at these seven sites ranged from +2.1°C to +6.1°C and mean annual precipitation ranged from 323 to 487 mm (Table 2). The virtual basin model of HEG catchments was run using daily precipitation data from the Adjusted and Homogenized Canadian Climate Data (AHCCD) (Mekis and Vincent, 2011; Vincent et al., 2012) collected at these seven locations. This dataset corrects shifts identified due to station relocation and changes in

265   observing practices and automation. Other discontinuities are adjusted with multiple linear regression using a penalized maximal t-test and a quantile-matching algorithm. For precipitation, corrections are applied to account for wind undercatch, evaporation, and gauge-specific wetting losses. Snowfall density corrections are derived based on coincident ruler and Nipher measurements. Trace precipitation is added. The daily precipitation data were converted

11

270 to hourly data required by CRHM using linear interpolation using the Observation module. The

other hourly forcing variables (temperature, relative humidity and wind speed) were taken from

Environment and Climate Change Canada observations for stations for the same seven locations.

The data were quality controlled and infilled using nearby station data.

275 Outputs from the baseline simulations (1965–2006) were used to assess whether the virtual HEG

basin model captured typical behaviour of real world HEG basins. The first five years of

simulations were excluded from the analysis to avoid the potential to misrepresent initial

conditions due to lag effects associated with antecedent conditions. Because the virtual basin is

designed to represent typical behaviour of a basin class, it does not reproduce the hydrology of

280 any specific basin and there is no direct physical analogue for which observations can be used to

completely assess performance. Previous studies have described the application of CRHM to

Canadian Prairie basins, and its ability to represent the region's predominant hydrological

processes is well established (Fang et al., 2010). These findings lend confidence that the virtual

basin model, informed by these other CRHM applications, can be applied for the diagnostic

285 purposes of this study. Furthermore, the aim of the simulations was not to simulate specific

basins in the region, but to assess the sensitivity of the hydrological regime to changes in climate

forcings. For this reason, spring snow water equivalent (SWE) values from snow courses and

mean annual hydrographs from hydrometric gauges at multiple sites within the HEG class were

compared to virtual basin model outputs to establish that the virtual basin model was capturing

290 the correct timing and magnitudes of important states and fluxes. The cluster median was not

used for comparison, as the range better demonstrates that model simulations are within the

envelope of plausible hydrological functioning as represented by snow on the ground and

12

streamflow, rather than being comparable to an individual basin within the class. Runoff ratios were also compared to establish if the virtual model was dividing the water budget in a

295 reasonable manner.

To evaluate spring snow accumulation, measured SWE at the Alberta Environment and Parks Wetaskiwin, Kneehill Valley and Innisfail East snow courses (Table 3) were compared against modelled SWE values for the cultivated HRU on the same day measurements were taken, using

300 climate data from the nearest station (Red Deer) (Table 2), for the period during which there was data overlap (1987–2006). It is recognized that snow accumulation in this region is not continuous throughout the winter as there can be significant ablation events, however, measurements are not taken earlier than late February. Mean monthly discharge depths from basins 100% within the HEG class were generated using climate data from four nearby

305 meteorological stations (Table 2) and were plotted and visually compared to the 14 Water Survey of Canada hydrometric stations gauging a stream within 100 km of one of the meteorological stations (Table 3). The selected basins and meteorological stations used for these evaluations were all in Alberta as this was where the class was most common and contiguous. These tests were to discern if the virtual basin model was capturing streamflow seasonality, variability and

310 annual runoff ratios.

After evaluation, the virtual basin model was used to explore (1) the role of climate variability within the HEG class, and (2) how hydrological patterns in the HEG class may change in response to potential future shifts in climate. Future climate scenarios were investigated using the

315 delta method, which applies uniform changes to historic daily air temperature and total

13

precipitation records (Lopez-Moreno et al., 2012; Rasouli et al., 2014). This method has the advantages of being computationally inexpensive, while avoiding bias, and preserving the covariances among variables, which are important in modeling cold-regions processes (Shook and Pomeroy, 2010). It has the disadvantage of not assessing sensitivity to changing extreme

320   events, and does not account for seasonal differences associated with climate change. Temperature increases of up to +6°C and precipitation decreases of 20% and increases of up to 30%, based on model projections for southern Alberta by 2050 (Zhang et al., 2018) were investigated. Within these ranges, multiple scenarios were run using increments of 1°C (temperature) and 10% (precipitation), totalling 35 scenarios for each of the seven climate

325   locations. For each 46 year run (1960 to 2006), the model takes around 7 minutes on a PC with CPU of Intel Core i7-8650U. The model could not be run for each of the 796 catchments in the HEG class as no adequate (i.e., low uncertainty, long period of record) gridded climate dataset was found. This limited the approach to running the model with the observations from seven stations. The 35 climate scenarios run for each of seven stations, took approximately 7 hours to

330   complete, running four simulations at a time on separate threads. These scenarios were used with the baseline runs to quantify sensitivity of snow accumulation and annual runoff to climate change. The spatial distributions of the sensitivities were mapped by extrapolating the values for these seven model locations (Table 2, Figure 2) in ArcGIS using ordinary kriging, assuming a spherical semivariogram and a lag value of 0.01964.

335

Results

The basin classification aligns well with previous and more recent classifications of Canadian Prairie basins based solely on streamflow regime (Ellis and Gray, 1966; Whitfield et al., 2021).

14

The HEG class occupies much of the western portion of the Canadian Prairies, and includes the

340   majority of southern Alberta and several isolated patches in both Saskatchewan and Manitoba

(Figure 2). Basins in this class tend to have a high fraction of native grasslands (40% of the

basin area). The elevations of these basins are amongst the highest in the Canadian Prairies. The

mean wetland density of 0.8 km^{-2} is amongst the smallest of the virtual basin classes (Wolfe et

al., 2019). Relatively dense drainage networks, coupled with the low wetland densities, result in

345   67% of the gross drainage area of a typical HEG basin contributing to runoff to the outlet in a

median flow year. This is a relatively high percentage for a Canadian Prairie basin.

Virtual basin model performance

Spring snow water equivalent

350   The virtual model captured the intra- and inter- annual variability in snow water equivalent

(SWE) quite well as indicated by comparison of CRHM HEG model simulated SWE (Red Deer

forcing data) and snow survey course data (Figure 4). The simulated values were calculated on

the same days as the snow surveys with dates varying between February and April. The

agreement is quite good considering that a) the forcing meteorological data were collected up to

355   60 km from the snow survey site, and b) that a virtual model, not tuned to the specific traits of

the snow survey location (e.g., vegetation height), was used. Simulated and observed means and

standard deviations were similar, but in two instances the model simulated SWE values notably

larger than observed during the comparison period (Figure 4). A best fit line with a slope of 1.1

and r^2 of 0.53 can be drawn among the simulated and observed data.

360

Runoff ratios and mean annual hydrographs

15

Runoff ratios were evaluated to determine if the virtual model was dividing the water budget correctly and producing a reasonable values of annual runoff. The mean simulated annual runoff ratios for HEG basins during 1965-2006 were 0.08 ± 0.11. These values are slightly lower than those from the literature. Most documented values of runoff ratios from this landscape are from hillslope and agricultural field scales. Woo and Rowsell (1993) estimated the mean value of annual runoff ratios from a grassland slope in Saskatchewan to be 0.13, which is within the range estimated by the model. Rainfall runoff ratios never exceeded 0.11 from grassland slopes observed by Neath and Chanasyk (1996) in the fescue grasslands of southern Alberta, which aligns well with the small discharges simulated by the model after spring melt (Figure 5). Pavlovskii et al. (2019) observed hillslope scale runoff ratios for snowmelt events over a wide range of values, with smaller values experienced in mid-winter (~0.3) (i.e., January and February) versus those in spring (~0.5).

The pattern of monthly discharge depths for the HEG model driven by climate forcings from the four Albertan climate locations were in good agreement with WSC gauged streams in the HEG classification (Figure 5). The monthly means were computed only from those years for which data were available for the HEG virtual model and for all gauging sites. As would be expected, the largest median monthly discharges are found in March through May, due to the spring melt of the accumulated snowpacks. The HEG virtual model did a good job of reproducing the timing and magnitudes of monthly discharge depths, particularly for March and April. This is impressive considering that the meteorological data were collected up to 60 km away from the stream gauging sites. The agreement between the simulated and gauged discharge depths was poorer in the summer, particularly in June and July. Mean conditions are well simulated, but

16

385   there are numerous instances of outliers in the observed time series, especially in May and June

(e.g., Lethbridge – Figure 5) not captured by the simulations. These events are undoubtedly the

result of small, fast-moving, short duration and intense convective rainfall events. The use of

daily precipitation reduces rainfall intensities in the meteorological forcings. Mean annual

runoff simulated over the HEG class was 10 mm, while that of the observed runoff from the

390   Water Survey of Canada gauges (Table 3) was 17 mm for the verification period of 1976 - 2006.

This bias of simulated discharge to small values aligns with the low runoff ratios and the poor

representation of convective rainfall events in the meteorological forcing discussed earlier.

Furthermore, the Water Survey of Canada often selects basins with contributing area fractions

that approach 1.0, as these are more efficient at producing streamflow and are easier to gauge

395   (Whitfield et al., 2020; Table 3). This differs from the contributing area fraction of 0.67 Wolfe et

al. (2019) measured as the median across the HEG class and used in the virtual model, and this

too could account for some of this low bias.

400   *Hydrological sensitivity to climate*

Snow cover

According to Li et al. (2019) and Zhang et al. (2018) future climates at the end of the 21st century

for the region could be as much as 30% wetter and 6°C warmer than the mid 20th century. The

effects of these changes on snow cover differ among HRUs (Figure 6; Table 4). The virtual

405   basin model results support the well-documented cold regions phenomenon in that the annual

peak SWE declines with warming (Najafi et al., 2017; Fang and Pomeroy, 2008). Source HRUs

for redistribution of snow (e.g., cultivated HRUs) experience small absolute changes. This may

17

be because blowing snow is affected by vegetation height, and a 6°C temperature rise does not

reduce the depth of accumulated snow below this threshold. HRUs receiving blowing snow,

410   such as wetlands, are more sensitive to warming (Figure 6). Warming of 6°C with no change in

mean annual precipitation decreased the simulated annual peak SWE by 27% in the cultivated

HRU and 51% in the wetland HRU in a climate comparable to Medicine Hat's– this is similar to

the sensitivity of snowdrift SWE to warming found by Fang and Pomeroy (2007; 2009) and is

due to the temperature sensitivity of the occurrence of blowing snow (Li and Pomeroy, 1997).

415

The date of annual peak SWE advances as annual air temperature warms (Figure 7). The

maximum advance in the date of the annual SWE peak (with no change in precipitation) is about

40 days. The date of peak SWE is more sensitive for climates typical of western Alberta (e.g.

Calgary) than those further east (e.g. Regina, Brandon). For example, an increase of 2°C is

420   needed to advance peak SWE date by 20 days near Calgary while a similar change requires 4°C

in Saskatchewan or Manitoba. This could be due to the colder initial temperatures further east

and north (Table 2). The nearly horizontal isochrones in Figure 7 suggest that the peak SWE

date is less a function of potential changes in precipitation amount and more a function of

anticipated temperature increases.

425

Warming is associated with an unequivocal decrease in the amount of snowfall, as the ratio of

rain to annual precipitation can increase by more than 10% with warming of 6°C (Figure 8).

Greater absolute decreases in wetland HRU SWE with warming were simulated using climate

from the eastern and northern edges of the class (i.e., Brandon, North Battleford and Regina)

430   (Figure 8). Under warming of 6°C annual peak SWE in the wetland HRU decreased by 100 mm

18

in a climate such as Medicine Hat's and as much as 220 mm in a climate such as Brandon's. The

absolute differences are due to the drier climates of the western locations (Table 2) that produce

smaller baseline annual peak SWE (157 mm at Medicine Hat vs. 409 mm at Brandon). The dry

climates also tended to be warmer, with a rain ratio more sensitive to warming (e.g., Lethbridge

435   vs. North Battleford, Figure 8). This results in a higher rate of loss in SWE in a climate such as

southern Alberta's (11% loss in annual peak SWE per °C) compared to climates from the eastern

portions (5% loss in annual peak SWE per °C) (Figure 9).

The changes in rainfall ratio illustrated in Figure 8 are evident when model simulations are

440   extrapolated across the HEG class (Figure 9). In much of the class, the annual maximum spring

SWE declines by more than 7% for each °C of warming. The most sensitive regions are in

western Alberta. When temperature change is held constant in the model, annual maximum SWE

changes almost proportionally with precipitation (Table 4). When the effects of temperature and

increasing precipitation are combined, there is a trend from east to west of larger decreases in

445   annual maximum SWE (Figure 9) with a warming and wetting climate. The predicted increase

of 30% in annual precipitation does not offset the impact of increases in temperature on SWE,

but the maximum annual SWE reductions per degree slow and range from 8% to 1%. The most

sensitive regions become concentrated in western Alberta as these become relatively drier.

450   *Streamflow*

Absolute decreases in runoff in the climates of Red Deer, Calgary and Lethbridge are all similar

under a warmer climate (Figure 10). Similar, but larger absolute decreases were simulated for the

colder climates of Brandon, Regina and North Battleford. These colder sites do not see as large

19

an increase in the rainfall ratio. (Figure 10). The smallest absolute change in annual runoff (7

455   mm) was found for areas with a Medicine Hat climate; however, this represents a 78% decrease

in runoff, as baseline runoff (9 mm; Table 4) was the smallest of the seven climates investigated.

Two reference conditions can result in lower relative change in runoff with warming, either a dry

or cold climate. This is evident in Figure 10. For each °C of warming, streamflow from HEG

basins with western Alberta climates decrease 15%, but in those with climates from the east and

460   north, this impact diminishes to decreases of 7% per °C of warming. Under a warmer and wetter

climate (6°C warming and 30% increase in annual precipitation) runoff in western portions still

experience decreases, but runoff in climates such as Brandon's increase and remain the same in

climates such as Saskatchewan's (Figure 11).

465   **Discussion**

The value of using a classification-based virtual basin approach is that it can be more efficient

than running models over each catchment in the region. Running the model with one

parameterization for a prototypical basin seven times for each climate record and with the 35

climate scenarios took seven hours. The subsequent kriging added another two hours for a total

470   of nine hours. Had there been reasonable climate data and all 796 basins in the HEG class were

run, the simulations would have taken 33 days.   A value of using a virtual basin model approach

rather than statistically extrapolating observed streamflow records and subsequently attempt to

disentangle controls on streamflow is the ability to isolate the influence of individual drivers

such as climate on hydrological responses. It can be difficult to glean from observations alone

475   how Canadian Prairie basins respond to change (Ehsanzadeh, 2016), because neither climate nor

land-use have remained stationary over the last half century. The model results imply that the

20

effects of temperature and precipitation changes will differ amongst HRUs. Increases in

precipitation may also cancel the direction and slow the magnitude of changes in streamflow

caused by increased temperatures, and these patterns vary across the region.

480

Hydrological sensitivity to climate change

The climate of the Canadian Prairie has been warming since the mid 20th century, with more

variable and greater precipitation (Gan, 1998; Millett et al., 2009), which has altered water

availability and flood characteristics (Pomeroy et al., 2009) which are of interest to water

485   managers and others. The uncertainty in future climate change projections, particularly in

precipitation (Bush and Lemmen, 2019), can be partially mitigated by the type of sensitivity

analysis conducted here, which can provide some information on how hydrological states and

fluxes may respond to alternate future climate conditions and to identify potential thresholds in

hydrological response. On the Canadian Prairie, the sensitivity to climate varies among

490   landscape components. There is confidence that annual warming of 6°C coupled with increases

in precipitation of 30% for HEG basins will induce detectable changes in spring snow condition,

with snowpacks in cultivated fields being less sensitive than wetlands. This type of change could

happen by the late 21st century (Bush and Lemmen, 2019). The sensitivity analysis implies that a

6°C warming alone could result in a 27% decline in annual maximum SWE in cultivated areas

495   and a 51% decline in depressions (Table 4). Blowing snow redistribution to depressions is

particularly sensitive to warming temperatures (Figure 8). The increase in temperature reduces

snow accumulation and, in turn, the redistribution of blowing snow to sink areas such as

depressions. Rasouli et al. (2014) also documented a non-linear response in annual peak SWE to

changes in temperature and precipitation, albeit in a mountain basin, where even a 30% increase

21

500   in precipitation could not compensate for the impacts of a 6°C warming on peak annual SWE.

The reduction in depression SWE has important implications for groundwater recharge as

depressions provide recharge at volumes disproportionate to their area on the landscape (Hayashi

and van der Kamp, 2005). Given that wetland depressions of the region are well known as

biodiversity hotspots (Mantyka-Pringle et al., 2019), these changes could have cascading effects

505   for biophysical systems.

Warming in the spring and reduced winter snow accumulation advance the timing of annual peak

SWE. With an increase of 6°C, annual peak SWE (with no change in precipitation) arrives 40

days earlier. This lengthening of the snow free season enhances the opportunity for evaporative

510   losses. The consequence for streamflow is a 44% decline in annual volume in a relatively dry

HEG climate such as Medicine Hat's (Table 4). Year-to-year variation may be more extreme

than suggested by the model because streamflow variability in the virtual basin model

simulations is less sensitive to inter-annual variation in climate than streamflow records for

gauged basins within the classification indicate. The results imply that changes in streamflow

515   due to warming can be offset by rising precipitation. Warming of this magnitude has occurred in

the last 60 years in southern Alberta (DeBeer et al., 2016) and the influence can be seen in the

smallest simulated SWE values typically appearing later in the simulation period (Figure 6).

Whitfield et al. (2020) found that basins that align with the HEG class are already responding to

this kind of climate change, exhibiting earlier and smaller runoffs, implying the area is becoming

520   drier. This is expected to continue with a further 2°C increase from current conditions (i.e., ~4 -

5°C from the mid 20th century) predicted by 2040 (Zhang et al., 2019). The value in

understanding the dynamics of combined impacts of temperature and precipitation changes on

22

SWE and streamflow is that it can highlight conditions that require more attention for mitigation. The diversity in streamflow response across the class, with greater decreases in streamflow

525 simulated in the west and possibly increases in the east (Figure 11) is similar to patterns observed by Johnson et al. (2005) across the entire Prairie Pothole Region over the last half of the 20th century. This kind of regional diversity may require a variety of policies and practices to be implemented across the region. This information could be helpful to water managers and policymakers in the region, including informing how small dams are operated (Muzik, 2001),

530 and wetlands are managed (Wilson et al., 2019).

The sensitivity of hydrological regimes documented here is similar to those of other studies conducted in the HEG classification and other nearby Prairie basins. Fang and Pomeroy (2007) and Pomeroy et al. (2007) attempted to determine the sensitivities of snow accumulation and

535 runoff to drought at Bad Lake, Saskatchewan. They adjusted initial soil moistures, vegetation heights and winter temperature and precipitation in ways that were informed by previous droughts and by the Global Climate Model predictions of the time. In their work, simulated snow cover duration decreased by 40 days given a 1°C winter warming and a 15% increase in winter precipitation, which may suggest greater sensitivity to changing winter climate than our

540 results under similar climate perturbations (single digit advancement of timing and modest decreases in amount of peak annual SWE; Figure 8). Muzik (2001) altered precipitation in a simulation of the Little Red Deer River, Alberta, and also found disproportionate changes in streamflow. Using downscaled values from Providing Regional Climates for Impacts Studies (PRECIS) and the Canadian Regional Climate Model (CRCM) and stochastic weather generator

545 data to force an application of the Soil Water Assessment Tool (SWAT) Model to a Prairie

23

Pothole basin in Saskatchewan, Zhang et al. (2011) found winter runoff increases to be a function of how frequently the air temperature exceeds 0°C. Using a similar approach, Muhammad et al (2018) show how warming can increase winter runoff, but this will not offset large decreases in spring and summer runoff. Simulations of Beaver Creek, Alberta to quantify

550   climate change impacts on snow accumulation and streamflow indicated that changes in precipitation of more than 10% are needed to alter annual runoff (Forbes et al., 2011). Rasouli et al. (2014) suggest that in mountainous areas of Alberta, runoff is more sensitive to changes in precipitation than temperature, however our results in the Canadian Prairies imply a more dynamic system. When there is warming and drying, runoff is sensitive to both precipitation and

555   temperature, but as conditions become wetter, the Prairie system becomes less sensitive to temperature. This is not necessarily due to changes in the snowpack, which responds in a more linear manner (Figure 9), but perhaps due to higher amounts of rainfall relative to the storage capacity of the landscape.

560   **Conclusions**

Virtual experiments have proven to be suitable to diagnose the hydrological response of basins in this landscape. This work introduces a modelling approach that can be used to generate new knowledge of hydrological behaviours at a regional scale under different boundary conditions. Where previous studies have focused on the sensitivity of individual basins in cold regions, and

565   conclusions about wider applicability were made by assuming the basins to be representative, this approach is meant to provide a methodology to assess how regional hydrology may respond to change by modelling a prototypical virtual basin that represents one type of basin in a region. The results of this study can be used, in part, to address demands for regional-scale information.

24

Such information is required to develop informed policies for climate change mitigation, but also

570 more broadly for water resources management in this landscape where land use is predominantly

for intensive agriculture. However, there are some limitations as the model assumes a

homogenous basin with an area of 100 km^2. It also does not account for second order changes

such as how agricultural practices will adapt in response to a warmer and wetter environment

over time and the role these landscape changes play in influencing hydrological processes.

575 Outputs of virtual experiments are less useful in predicting exact future system states than in

specifying how alternative climate possibilities would alter hydrological behaviour. Therefore,

the model output should be interpreted with some caution as the approach is less useful in

predicting exact future system states than in specifying how alternative choices could alter the

tendency to move towards each of those conditions.

580

The simulations provide novel insights into the interactions that influence the hydrological

response of the High Elevation Grasslands classification to different climates. The experimental

design was successful in simulating the sensitivity of a basin class in the Canadian Prairie to

expected climate change. The scenario results show that the hydrology of HEG basins in the

585 Canadian Prairies is highly sensitive to changes in climate. What should be considered is that the

changes simulated by the virtual basin model are already underway. For instance, the mean

annual temperature at Medicine Hat for the baseline simulation period (1960 – 2006) was 5.5°C.

The region is expected to warm 2°C above the 1976 – 2005 climate normal (6.1°C) to 8.1°C by

2040 (Zhang et al., 2019). It is reasonable to expect that the 2°C warming scenario presented

590 here will be the normal by the end of this decade. Water managers should expect 38% less snow

in depressions and 20% less snow in cultivated fields in the spring, and 33% less annual runoff.

25

Of course, this depends on how wet the next ten years are. Simulations imply that a 10-20%
increase in average annual precipitation would be needed to offset the warming. These are
profound differences in the response of the hydrological regime of the region to a warmer and
595 wetter climate. Water managers and agricultural producers need to consider carefully how they
can adapt their practices in light of such changes.

References

Armstrong, R. N., Pomeroy, J. W. and Martz, L. W. Evaluation of three evaporation estimation methods in a Canadian prairie landscape. Hydrol. Process, 22, 2801-2815, 2008.

Armstrong R.N, Pomeroy J.W. and Martz L.W. Variability in evaporation across the Canadian Prairie region during drought and non-drought periods. J. Hydrol., 521, 182-195, 2015.

Ayers, H. D. Influence of soil profile and vegetation characteristics on net rainfall supply to runoff. In: Proceedings of Hydrology Symposium V. 1, 198-205, 1959.

Brannen, R., Spence, C., and Ireson, A. Influence of shallow groundwater-surface water interactions on the hydrological connectivity and water budget of a wetland complex. Hydrol. Process., 29, 3862–3877, 2015.

Bush, E. and Lemmen, D.S., (2019): Canada's Changing Climate Report; Government of Canada, Ottawa, ON. 444 p.

DeBeer, C. M., Wheater, H. S., Carey, S. K., and Chun, K. P. Recent climatic, cryospheric, and hydrological changes over the interior of western Canada: a review and synthesis, Hydro. Earth Syst. Sci., 20, 1573-1598, 2018

Di Giammarco, P., Todini, E. and Lamberti, P. A conservative finite elements approach to overland flow: the control volume finite element formulation. J. of Hydrol., 175, 267-291, 1996.

Dumanski, S., Pomeroy, J. W. and Westbrook, C. J. Hydrological regime changes in a Canadian Prairie basin, Hydrol. Process., 29, 3893-3904, 2015.

Dunn, S. M., McDonnell, J. J. and Vaché, K. B. Factors influencing the residence time of catchment waters: A virtual experiment approach. Water Res. Res., 43, 2007.

Ehsanzadeh, E., Spence, C., van der Kamp, G. and McConkey, B., On the behaviour of dynamic contributing areas and flood frequency cures in North American Prairie watersheds. J. Hydrol., 414-415, 364-373, 2012.

Ehsanzadeh, E. Impact of climate variability and wetland drainage on watershed response in depression dominated landscapes. Int. J. River basin Mgmt., 16, 169-178, 2016.

Ellis, W.H., and Gray, D.M. Interrelationships between the peak instantaneous and average daily discharges of small prairie streams. Canadian Agricultural Engineering, 8,1-3, 1966.

Ellis C.R., Pomeroy J.W., Brown T. and MacDonald J.P. Simulation of snow accumulation and melt in needleleaf forest environments. Hydrol. Earth Syst. Sci., 14, 925-940. 2010.

Fang, X. and Pomeroy, J. W. Snowmelt runoff sensitivity analysis to drought on the Canadian prairies. Hydrol. Process., 21, 2594-2609, 2007.

27

Fang, X. and Pomeroy, J. W. Drought impacts on Canadian prairie wetland snow hydrology. Hydrol. Process., 22, 2858-2873, 2008.

660   Fang, X. and Pomeroy, J. W. Modelling blowing snow redistribution to prairie wetlands, Hydrol. Process., 23, 2557-2569, 2009.

Fang, X., Pomeroy, J. W., Westbrook, C. J., Guo, X., Minke, A. G. and Brown, T. Prediction of snowmelt derived streamflow in a wetland dominated prairie basin, Hydrol. Earth Syst. Sci., 14, 991-1006, 2010.

665   Forbes, K. A., Kienzle, S. W., Coburn, C. A., Byrne, J. M. and Rasmussen, J. Simulating the hydrological response to predicted climate change on a watershed in southern Alberta, Canada. Clim. Change, 105, 555-576, 2011.

670   Gan, T. Y. Hydroclimatic trends and possible climatic warming in the Canadian Prairies. Water Res. Res., 34, 3009-3015, 1998.

Godwin, R. B. and Martin, F. R.J. Calculation of gross and effective drainage areas for the Prairie Provinces. In: Canadian Hydrology Symposium - 1975 Proceedings, 11-14
675   August 1975, Winnipeg, Manitoba. Associate Committee on Hydrology, National Research Council of Canada, 219-223, 1975.

Gray, D. M. and Landine, P. G. An energy-budget snowmelt model for the Canadian Prairies. Can. J. Earth Sci., 25, pp.1292-1303. 1988.

680   Gray, D. M., Landine, P. G. and Granger, R. J. Simulating infiltration into frozen prairie soils in streamflow models. Can. J. Earth Sci., 22, 464-472, 1985.

Haque, A., Ali, G., Badiou, P. Hydrological dynamics of prairie pothole wetlands: Dominant
685   processes and landscape controls under contrasted conditions, Hydrol. Proc., 32, 2405-2422, 2017.

Hayashi, M. and van der Kamp, G. Snowmelt contribution to groundwater recharge. In: The Science, Impacts and Monitoring of Drought in Western Canada, Sauchyn, D. et al.,
690   (eds.), Great Plains Research Centre, Regina, 2005.

Horn, A. L., Hörmann, G. and Fohrer, N. Application of a virtual watershed in academic education. Adv. in Geosci., 5, 137-141, 2005.

695   Johnson, W.C., Millett, B.V., Gilmanov, T., Voldseth, R.A., Guntenspergen, G.R., and Naugle, D.E. Vulnerability of northern prairie wetlands to climate change. Bioscience, 55, 863-872, 2005

Knoben, W. J. M., R. A. Woods, and Freer, J.E. A Quantitative Hydrological Climate
700   Classification Evaluated With Independent Streamflow Data. Water Res. Res., 54, 5088-5109, 2018.

Lehner, B. and Grill, G. Global river hydrography and network routing: baseline data and new approaches to study the world's large river systems. Hydrol. Process., 27, 2171-2186, 2013.

Li L. and Pomeroy J.W. Probability of occurrence of blowing snow. J. Geophys. Res., 102, 21955-21964, 1997.

Li, Y., Li, Z., Zhang, Z., Chen, L., Kurkute, S., Scaff, L. and Pan, X. High-Resolution Regional Climate Modeling and Projection over 5 Western Canada using a Weather Research Forecasting Model with a Pseudo-Global Warming Approach. Hydrol. Earth Syst. Sci. Discuss., 2019.

López-Moreno J. I., Pomeroy J. W., Revuelto J., Vicente-Serrano S. M. Response of snow processes to climate change: spatial variability in a small basin in the Spanish Pyrenees. Hydrol. Process., 27, 2637–2650, 2012.

Lopez-Moreno J. I., Pomeroy J. W., Alnso-González E., Morán-Tejeda, Revuelto, J. Decoupling of warming mountain snowpacks from hydrological regimes. Env. Res. Lett., 15, 114006, 2020

Male, D. H., and Gray, D.M.. Snowcover ablation and runoff. In Handbook of snow: principles, processes, management and use, Gray, D.M. and Male, D.H. (eds.), Pergamon Press, Toronto, pp. 776, 1981

Mallard, J., McGlynn, B. and Covino, T. Lateral inflows, stream-groundwater exchange, and network geometry influence stream water composition. Water Res. Res., 50, 4603-4623, 2014.

Mantyka-Pringle, C., Leston, L., Messmer, D., Asong, E., Bayne, E. M., Bortolotti, L. E., Sekulic, G., Wheater, H., Howerter, D. W., and Clark, R. G. 2019. Antagonistic, synergistic and direct effects of land use and climate on prairie wetland ecosystems: Ghosts of the past or present? Diversity and Distributions 25: 1924–1940. doi:10.1111/ddi.12990.

McDonnell, J. J. and Woods, R. On the need for catchment classification. J. of Hydrol., 299, 2-3. 2004.

Mekis, É. and Vincent, L. A. An overview of the second generation adjusted daily precipitation dataset for trend analysis in Canada. Atmosphere-Ocean, 49, 163-177, 2011.

Millett, B., Johnson, W. C. and Guntenspergen, G. Climate trends of the North American prairie pothole region 1906–2000. Clim. Change, 93, 243-267, 2009.

Monteith, J. L. Evaporation and environment. In: Symposia of the Society for Experimental Biology, V. 19, 205-234, Cambridge University Press, Cambridge, 1965.

Morton, F. I. Operational estimates of lake evaporation. J. of Hydrol., 66, 77-100, 1983.

750

Muhammad, A., Evenson, G.R., Stadnyk, T.A., Boluwade, A., Jha, S.K., and Coulibaly, P. Assessing the importance of potholes in the Canadian Prairie Region under future climate change scenarios. Water, 10, 1657, doi:10.3390/w10111657.

755 Muzik, I. Sensitivity of hydrologic systems to climate change. Can. Water Res. J., 26, 233-252, 2001.

Najafi, M. R., Zwiers, F. and Gillett, N. Attribution of the observed spring snowpack decline in British Columbia to anthropogenic climate change. Journal of Climate, 30, 4113-4130, 760 2017.

Naeth, M. A. and Chanasyk, D. S. Runoff and sediment yield under grazing in foothills fescue grasslands of Alberta. Journal of the American Water Resources Association, 32, 89-95, 1996.

765

Pavlovskii, I., Hayashi, M., and Itenfisu, D. Midwinter melts in the Canadian prairies : energy balance and hydrological effects. Hydro. Earth Syst. Sci., 23, 1867-1883, 2019.

Pomeroy, J. W., de Boer, D., and Martz, L. W. Hydrology and Water Resources of 770 Saskatchewan, Centre for Hydrology Report No. 1, University of Saskatchewan, Saskatoon, 25 pp., 2005.

Pomeroy, J. W., Gray, D. M., Brown T., Hedstrom, N. H., Quinton, W. L., Granger, R.J., Carey, S. K. The cold regions hydrological model: a platform for basing process representation 775 and model structure on physical evidence. Hydrol. Process., 21, 2650–2667, 2007.

Pomeroy, J., Fang, X., Shook, K., Westbrook, C. and Brown, T. Informing the Vermilion River watershed plan through application of the Cold Regions Hydrological Model platform. Centre for Hydrology Report No. 12, University of Saskatchewan, Saskatoon, 155 pp., 780 2012

Pomeroy, J. W., Fang, X. and Williams, B. Impacts of Climate Change on Saskatchewan's Water Resources. Centre for Hydrology Report No. 6, University of Saskatchewan, Saskatoon, 46 pp., 2009.

785

Pomeroy, J. W., Fang, X., Westbrook, C., Minke, A., Guo, X. and Brown, T. Prairie Hydrological Model Study Final Report. Centre for Hydrology Report No. 7, University of Saskatchewan, Saskatoon, 113 pp., 2010.

790 Pomeroy, J. W., Granger, R., Pietroniro, A., Elliott, J., Toth, B. and Hedstrom, N. Classification of the boreal forest for hydrological processes. In: Proceedings of the Ninth International Boreal Forest Research Association Conference, Woxholtt, S. (ed.), 49-59, 1999.

30

Pomeroy, J. W. and Li, L. Prairie and arctic areal snow cover mass balance using a blowing snow model. Journal of Geophysical Research: Atmospheres, 105(D21), 26619-26634, 2000.

Priestley, C. H. B. and Taylor, R. J. On the assessment of surface heat flux and evaporation using large-scale parameters. Monthly Weather Review, 100, 81-92, 1972.

Rasouli, K., Pomeroy, J W., Janowicz, J. R., Carey, S. K., and Williams, T. J. Hydrological sensitivity of a northern mountain basin to climate change. Hydrological Processes, 28, 4191-4208, 2014.

Seo, Y. and Schmidt, A.R. Network configuration and hydrograph sensitivity to storm kinematics. Water Resources Research, 49, 1812-1827, 2013.

Shaw, D.A., Vanderkamp, G., Conly, F.M., Pietroniro, A. and Martz, L. The fill–spill hydrology of prairie wetland complexes during drought and deluge. Hydrological Processes, 26, 3147-3156, 2012.

Shook, K.R. and Pomeroy, J.W. Hydrological effects of the temporal variability of the multiscaling of snowfall on the Canadian prairies. Hydrology and Earth System Sciences, 14, 1195-1203, 2010.

Shook, K.R. and Pomeroy, J.W. Memory effects of depressional storage in Northern Prairie hydrology, 3890 – 3898, 2011.

Shook, K.R. and Pomeroy, J.W. Changes in the hydrological character of rainfall on the Canadian prairies. Hydrological Processes, 26, 1752-1766, 2012.

Shook, K.R., Pomeroy, J.W., Spence, C. and Boychuk, L. Storage dynamics simulations in prairie wetland hydrology models: evaluation and parameterization. Hydrol. Proc., 27, 1875-1889, 2013

Sicart, J.E., Essery, R.L., Pomeroy, J.W., Hardy, J., Link, T. and Marks, D. A sensitivity study of daytime net radiation during snowmelt to forest canopy and atmospheric conditions. Journal of Hydrometeorology, 5, 774-784, 2004.

Spence, C. and Mengistu, S.G. On the relationship between flood and contributing area. Hydrological Processes, 33, 1980-1992, 2019.

Spence, C., Wolfe, J.D., Whitfield, C.J., Baulch, H.M., Basu, N.B., Bedard-Haughn, A.K., Belcher, K.W., Clark, R.G., Ferguson, G.A., Hayashi, M., Liber, K., McDonnell, J.J., Morrisey, C.A., Pomeroy, J.W., Reed, M.G. and Strickert, G. Prairie water: a global water futures project to enhance the resilience of prairie communities through sustainable water management. Canadian Water Resources Journal, 44, 115-126. 2019

Stichling, W. and Blackwell, S.R. Drainage area as a hydrologic factor on the Canadian prairies, IUGG Proceedings, Toronto, Ontario, 1957.

Tiner, R.W. Geographically isolated wetlands of the United States. Wetlands, 23, 494-516, 2003.

van Meter, K.J. and Basu, N. Signatures of human impact: size distribution and spatial organization of wetlands in the Prairie Pothole landscape. Ecological Applications, 25, 451-465, 2015.

Vincent, L.A., Wang, X.L., Milewska, E.J., Wan, H., Yang, F., and Swail, V. A second generation of homogenized Canadian monthly surface air temperature for climate trend analysis, Journal of Geophysical Research, 117, D18110, doi:10.1029/2012JD017859, 2012

Wagener, T., Sivapalan, M., Troch, P. and Woods, R. Catchment classification and hydrologic similarity. Geography Compass, 1, 901-931, 2007.

Walmsley, J.L., Taylor, P.A. and Salmon, J.R. Simple guidelines for estimating windspeed variations due to small-scale topographic features–an update. Climatological Bulletin, 23, 3-14. 1989.

Weiler, M. and McDonnell, J.J. Virtual experiments: a new approach for improving process conceptualization in hillslope hydrology. Journal of Hydrology, 285, 3-18, 2004.

Weiler, M. and McDonnell, J. J. Testing nutrient flushing hypotheses at the hillslope scale: A virtual experiment approach. Journal of Hydrology, 319, 339-356, 2006.

Whitfield, P.H., Kraaijenbrink, P.D.A., Shook, K.R. and Pomeroy, J.W. The spatial extent of hydrological and landscape changes across the mountains and prairies in the Mackenzie and Nelson River basins based on data from the warm-season time window. Hydrol. Earth Syst. Sci., 25, 2513-2541, 2021

Whitfield, P.H., Shook, K.R., and Pomeroy, J.W. Spatial patterns of temporal changes in Canadian Prairie streamflow using an alternative trend assessment approach. Journal of Hydrology, 582, 124541, 2020.

Wilson, H. F., Casson, N. J., Glenn, A. J., Badiou, P., and Boychuk, L. Landscape controls on nutrient expot during snowmelt and an extreme rainfall runoff event in northen agricultural watersheds. Journal of Environmental Quality, 48, 841-849, 2019.

Wolfe, J. D., Shook, K.R., Spence, C. and Whitfield, C.J., 2019. A watershed classification approach that looks beyond hydrology: application to a semi-arid, agricultural region in Canada. Hydrology and Earth System Sciences, 23(9), pp.3945-3967.

Woo, M. K. and Rowsell, R. D. Hydrology of a prairie slough. Journal of Hydrology, 146, 175-207, 1993.

885

Zhang, X., Flato, G., Kirchmeier-Young, M.,Vincent, L., Wan, H., Wang, X., Rong, R., Fyfe, J., Li, G., Kharin, V.V. Changes in Temperature and Precipitation Across Canada; Chapter 4 in Bush, E. and Lemmen, D.S. (Eds.) Canada's Changing Climate Report. Government of Canada, Ottawa, Ontario, 112-193, 2019

890

Zhang, H., Huang, G.H., Wang, D. and Zhang, X. Uncertainty assessment of climate change impacts on the hydrology of small prairie wetlands. Journal of hydrology, 396, 94-103, 2011.

895

Zhang, B., Schwartz, F.W., and Liu, G. Systematics in the size structure of prairie pothole lakes through drought and deluge. Water Res. Res., 45, 1–12, 2009.

TABLES

Table 1: Cold Regions Hydrological Model parameters for the High Elevation Grassland virtual basin model. The suffix "-w" in the hydrologic response unit (HRU) name indicates HRUs in the wetland catena. LAI denotes leaf area index.

HRU	Fraction of basin	LAI	Fetch (m)	Vegetation height (m)	Stalk density ($/m^2$)	Stalk diameter (m)	Manning's n
Channel (CH)	0.01	0.001	300	0.50	1	0.003	0.07
Cultivated (CL)	0.32	0.001	1000	0.20	320	0.003	0.17
Cultivated (CL-w)	0.13	0.001	1000	0.20	320	0.003	0.17
Fallow (FL)	0.004	0.001	1000	0.01	320	0.003	0.05
Fallow (FL-w)	0.002	0.001	1000	0.01	320	0.003	0.05
Grassland (GL)	0.30	0.001	500	0.40	320	0.003	0.2
Grassland (GL-w)	0.12	0.001	500	0.40	320	0.003	0.2
Shrubland (SL)	0.02	0.001	300	1.50	100	0.01	0.2
Shrubland (SL-w)	0.006	0.001	300	1.50	100	0.01	0.2
Woodland (WL)	0.02	0.4	300	6.00	100	0.1	0.4
Woodland (WL-w)	0.006	0.4	300	6.00	100	0.1	0.4
Wetland (WT)	0.07	0.001	300	1.50	1	0.01	0.2

Albedo

Bare ground	0.17
Snow (fresh)	0.85

Wetland area Generalized Pareto Distribution parameters

Shape (ξ)	1.152
Scale (β)	2070.62

Table 2: Climate characteristics (1981–2010 climate normal) of the seven selected locations located in and near the High Elevation Grassland class. T_a is mean annual temperature and P denotes mean annual precipitation

Location	Latitude	Longitude	T_a ($^{\circ}$C)	P (mm)
Red Deer, Alberta	52° 16' 22" N	113° 48' 36" W	2.8	487
Calgary, Alberta	51° 02' 37" N	114° 04' 33" W	4.4	419
Medicine Hat, Alberta	50° 02' 43" N	110° 40' 05" W	6.1	323
Lethbridge, Alberta	49° 41' 32" N	112° 51' 07" W	5.9	380
North Battleford, Saskatchewan	52° 46' 37" N	108° 17' 43" W	2.1	374
Regina, Saskatchewan	50° 26' 58" N	104° 37' 10" W	3.1	390
Brandon, Manitoba	49° 50' 51" N	99° 56' 50" W	2.2	474

Table 3: Sources of observed data for model evaluation. Water Survey of Canada's (WSC) hydrometric data were obtained from the HYDAT database, available at https://collaboration.cmc.ec.gc.ca/cmc/hydrometrics/www/. Effective drainage area is defined by Godwin and Martin (1975) as the drainage area that contributes streamflow to the gauged location during the median annual flood was also obtained from HYDAT. The snow water equivalent data were obtained from Alberta Environment and Parks snow courses.

Station name	Period of record used for validation	Associated climate location(s)	Gross drainage area (km^2)	Effective drainage area (km^2)
Hydrometric				
Battle River near Ponoka (05FA001)	1976 - 2006	Red Deer	1820	1550
Maskwa Creek No. 1 above Bearhills Lake (05FA014)	1976 - 2006	Red Deer / Calgary	79.1	61.2
Renwick Creek near Three Hills (05CE011)	1976 - 2006	Red Deer / Calgary	58.9	58.1
Ray Creek near Innisfail (05CE010)	1976 - 2006	Red Deer	44.4	44.4
Blindman River near Blackfalds (05CC001)	1976 - 2006	Red Deer	1796	1460
Battle Creek at Alberta Boundary (11AB117)	1976 - 2006	Medicine Hat	111	111
Gros Ventre Creek near Dunmore (05AH037)	1976 - 2006	Medicine Hat	215	206
Beaver Creek near Brocket (05AB013)	1976 - 2006	Lethbridge	257	257
Prairie Blood Coulee near Lethbridge (05AD035)	1976 – 2006	Lethbridge	214	214
Pothole Creek near Magrath (05AE011)	1976 – 2006	Lethbridge	372	351
Snake Creek near Vulcan (05AC030)	1976 – 2006	Lethbridge	350	346
Trout Creek near Granum (05AB005)	1976 – 2006	Lethbridge	441	441
West Arrowwood Creek near Ensign (05BM018)	1976 – 2006	Calgary	30	30
Snow Course				
Innisfail East (05CE801)	1987 - 2006	Red Deer		
Wetaskiwin (05FA801)	1987 - 2006	Red Deer		
Kneehill Valley	1987 - 2006	Red Deer		

Table 4: Simulated annual peak snow water equivalent (SWE) for different hydrologic response units (HRU) and basin runoff under different climate scenarios (temperature and precipitation

expressed as °C or % change from reference scenario, respectively) with a climate similar to that at Medicine Hat.

Variable	HRU	Temperature: 0 ºC Precipitation (%)					Temperature: 2 ºC Precipitation (%)					Temperature: 6 ºC Precipitation (%)				
		0	−20	+10	+20	+30	0	−20	+10	+20	+30	0	−20	+10	+20	+30
Annual peak SWE (mm)	Cultivated	60	46	68	77	84	48	38	52	59	65	44	34	49	54	59
	Grassland	67	50	74	84	92	50	40	56	64	70	46	35	51	57	62
	Shrubland	71	52	80	92	101	51	40	58	67	75	46	35	52	59	65
	Wetland	133	93	154	178	197	82	54	98	120	138	65	44	77	94	111
Annual runoff (mm)		9	4	12	15	19	6	3	8	11	15	5	2	6	9	12

930

FIGURE CAPTIONS

935

Figure 1: Components of the classification-based virtual basin modeling platform.

Figure 2: Basin classification map for the Prairie ecozone in western Canada. Colours represent the seven basin classes and excluded areas (lakes and urban areas) are shown in white. The focus of this paper is on the High Elevation Grasslands, shown in green. Climate data from the cities noted in this figure were used to drive the virtual model. Boundaries of the basins used in the classification are shown in grey

940

Figure 3: Illustration of HRU distribution in the virtual model of a typical High Elevation Grasslands basin. The illustration depicts two catenas, one with a wetland complex and one without, separated by the dashed line. Relative areas are meant to approximate those listed in Table 1. SL denotes shrublands; WL, woodlands; and CH a channel HRU (see Table 1). The fallow and shrubland and woodland HRU's directed through the wetland complex are too small to be shown at this scale.

945

950

Figure 4: Comparison of 1987–2006 simulated SWE using the Red Deer climate forcings and SWE observed at the three Alberta Environment and Parks (AEP) snow courses near Red Deer.

Figure 5: Simulated (CRHM) monthly simulated discharge depths (red) using forcing from four of the climate locations compared with those from WSC gauged streams (blue). The horizontal line in the middle of the box denotes the mean, and the top and bottom of the box denote plus or minus one standard deviation. The whiskers denote 10% and 90% percentiles. Circles represent values beyond these percentiles.

955

Figure 6: Peak virtual model simulated annual snow accumulation (SWE) for cultivated, grassland, shrubland, and wetland HRUs for Medicine Hat climate under baseline conditions (1965–2006: 0°C) and warming of air temperature up to 6°C.

960

Figure 7: Change in peak annual snow water equivalent date at the wetland HRU under warming and changes in precipitation for simulations using the perturbed 30 year climate data. Negative

965

signs in peak SWE date plot represent advance in time and positive signs denote delay in date of peak SWE accumulation.

Figure 8: Changes in mean annual peak snow water equivalent (mm) in a wetland HRU, annual blowing snow sublimation (mm) and the change in the mean ratio of rain in annual precipitation (%). Change of rain ratio was calculated by rain ratio of scenario (%) minus the rain ratio of reference (%).

Figure 9: Sensitivity of annual maximum snow water equivalent (SWE) in the grassland HRU to warming, assuming no change in precipitation (top), and a 30% increase in mean annual precipitation (bottom).

Figure 10: Change in runoff (mm), annual blowing snow sublimation (mm) and change in the ratio of rain in annual precipitation (%) for each climate (annual blowing snow sublimation and ratio of rain as in Figure 8).

Figure 11: Sensitivity of annual runoff from HEG basins to warming assuming no change in precipitation (top), and a 30% increase in mean annual precipitation (bottom). Note that the black bar on the legend for the bottom panel denotes the colour at which there is no change.