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Snowmelt runoff sensitivity analysis to drought on the Canadian prairies

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Abstract:

The Canadian prairies are subject to severe extended droughts that are characterized by warmer temperatures, lower precipitation, lower soil moisture and sparser vegetation than normal conditions. The physically based cold regions hydrological modelling platform (CRHM) provides a possible means to analyse the sensitivity of prairie snowmelt processes to drought. The model was tested against detailed observations from Creighton Tributary of the Bad Lake Research Basin, Saskatchewan for the 1974–1975 and 1981–1982 hydrological years and found to perform satisfactorily in reproducing snow accumulation and streamflow without parameter calibration. By lowering winter precipitation and raising winter air temperature from actual meteorological observations and by lowering fall soil moisture and vegetation height parameters, the resulting drought condition sensitivity of snow accumulation, snow cover duration, sublimation of blowing snow, evaporation, infiltration into frozen soils, soil moisture storage change, snowmelt runoff and streamflow discharge was estimated. Snow accumulation and snow cover duration were relatively insensitive to meteorological changes associated with drought because the suppression of blowing snow sublimation moderated reduced snowfall. Infiltration, soil moisture storage change and evaporation were also relatively insensitive to drought conditions. However, lower precipitation, higher air temperature and lower initial soil moisture caused a marked reduction in snowmelt runoff. Similarly, large reductions in streamflow discharge were caused by diminished winter precipitation, increased winter air temperature and decreased fall soil moisture content. A scenario showed that a combination of these factors could cause complete cessation of spring streamflow even under moderate drought of 15% reduction in winter precipitation and 2·5 °C increase in winter mean air temperature. Results show that spring runoff and streamflow discharge are inherently unstable in the Canadian prairie environment, and so, magnify the impacts of drought, and through multi-season storage and vegetation change can cause the impacts of hydrological drought to persist for several seasons after meteorological drought has ended. Copyright © 2007 John Wiley & Sons, Ltd.

KEY WORDS prairie runoff; snowmelt; snow accumulation; blowing snow; modelling; infiltration to frozen soil; drought; climate impacts; Cold Regions Hydrological Model

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INTRODUCTION

Droughts are frequent on the Canadian prairies. Over half the years of three decades, 1910–1920, 1930–1939, and 1980–1989 were in drought (Nkemdirim and Weber, 1999) and the recent drought of 1999–2004 was the most severe on record in parts of the prairies (Bonsal and Wheaton, 2005). Drought has important hydrological implications on the prairies as both snowfall and soil moisture are greatly reduced and this may impact streamflow and replenishment of water bodies (Nkemdirim and Weber, 1999). Wheaton *et al*. (2005) reported exceptionally low precipitation and low snow cover in the winter of 2000–2001, with the greatest anomalies of precipitation in Alberta and western Saskatchewan along with near-normal temperature in most of southern Canada. The reduced snowfall led to lower snow accumulation. A loss in agricultural production over Canada by an estimated $3·6 billion in 2001–2002 was attributed to this drought (Wheaton *et al*., 2005).

Drought studies have traditionally emphasized increased temperature and decreased precipitation as fundamental characteristics of drought (Shabbar and Khandekar, 1996). However, drought dynamics are described by more than precipitation and temperature as water supply is strongly affected by complex drought characteristics in the Canadian prairie environment. Despite snowfall comprising only one-third of annual precipitation, over 80% of annual runoff is derived from snowmelt in this environment (Gray *et al*., 1989). The accumulation and melt of snow is, therefore, of primary importance in controlling prairie runoff generation (Norum *et al*., 1976). Because snow is redistributed, the relationship between snow accumulation and snowfall is not straightforward, and blowing snow processes and their interaction with the surface must be taken into account to calculate snow accumulation (Pomeroy *et al*., 1993). For instance, vegetation height plays an important role in controlling amount of snow redistribution by wind (Pomeroy and Gray, 1994, 1995). Variations in vegetation height can affect surface roughness such that wind speeds are changed; this, in turn, has an influence on the wind redistribution of snow. As the stubble height increases from 1 to 40 cm on agricultural fields nearby

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Regina, the loss of snow accumulation to blowing snow decreases by about 22% of the mean seasonal snowfall (Pomeroy and Gray, 1995). The amount of snow accumulation eventually controls both infiltration to frozen soils and snowmelt runoff, so the effect of differences in vegetation height on snow accumulation can be reflected in snowmelt runoff. Snowmelt infiltration into unsaturated frozen soils (INF) is directly affected by snow accumulation (SWE) as demonstrated by Gray's expression for snowmelt infiltration (Gray *et al*., 1985, 1986):

$$\mathrm{INF} = 5 \cdot (1 - S_I) \cdot \mathrm{SWE}^{0\cdot584} \quad (1)$$

where SWE is snow water equivalent (SWE) (mm) and S_{I} is the initial soil saturation of 0–300 mm soil layer (dimensionless). An inverse relation between snowmelt infiltration and initial soil saturation is specified by Equation (1). Thus, when initial soil saturation decreases under drought conditions, snowmelt infiltration is expected to rise. Soil saturation can be found from the volumetric water content and soil porosity.

Adjusting snowfall and air temperature from actual meteorological data and changing model parameters such as initial soil moisture and vegetation height can create scenario conditions that typify drought. Droughts are warmer and drier than non-drought periods (Maybank *et al*., 1995; Wheaton *et al*., 2005). Though there are few analyses of the winter temperatures, Nkemdirim and Weber (1999) reported that winter precipitation was 68–73% of normal in 1981, 1983, 1984, 1985, and 1986; it was 43 and 46% of the normal in 1988 and 1989, respectively during a severe drought. Vegetation is also sparser and shorter during multi-year droughts as farmers have fewer crops to leave as standing stubble and as native grasses, bushes and trees die back, becoming sparse and short. The recovery time for native vegetation is longer than that for cropped vegetation in that it can take several years for bush and tree dieback to be replaced with new growth, whereas one normal summer's rainfall can restore crops heights. Such drought scenarios could be applied with a physically based hydrological model to estimate the hydrological sensitivity to climate variability typified by drought. As a result, the sensitivity of snowmelt runoff to drought can be examined.

The physically based cold regions hydrological modelling platform (CRHM) provides a means to analyse the sensitivity of prairie snowmelt runoff to drought. CRHM is based on a modular, object-oriented structure in which component modules represent basin descriptions, observations, or algorithms for calculating hydrological processes (Pomeroy *et al*., 2007). These hydrological processes include wind redistribution of snow, snowmelt, infiltration into unsaturated frozen soils, and snowmelt runoff. The corresponding modules are the Prairie Blowing Snow Model (Pomeroy *et al*., 1993; Pomeroy and Li, 2000), the Energy–Budget Snowmelt Model (Gray and Landine, 1988), Gray's equation for prairie snowmelt infiltration (Gray *et al*., 1985), and Clark's lag and route surface runoff routing (Clark, 1945). To calculate the water balance of a basin, modules are linked into a purpose built model for the basin of interest. Basins are composed of a number of hydrological response units (HRUs). HRUs are landscape units sharing the same properties with respect to the hydrological processes, in that they can be described by unique sets of parameters, variables and fluxes (Pomeroy *et al*., 2007). For each HRU, simulations can estimate snow accumulation, melt rate, cumulative snowmelt infiltration, and the amount of runoff from snowmelt. For the simulation period, the basin mean snow accumulation, infiltration and snowmelt discharge can be estimated.

The objectives of this study are to

1. Evaluate the suitability of the Cold Regions Hydrological Model for Canadian prairie hydrology simulations during winter and snowmelt periods including drought.
2. Determine the sensitivity of snow accumulation, infiltration, water storage and snowmelt runoff to drought conditions.
3. Describe the hydrological drought characteristics that ensue from a scenario progression of a prairie drought as a small catchment enters and leaves a drought condition.

STUDY AREA AND FIELD OBSERVATIONS

The study was conducted in Creighton Tributary of Bad Lake International Hydrological Decade (IHD) Research Basin (Figure 1). Bad Lake (51°23′N, 108°26′W, 650 m a.s.l.) is an internally drained basin near Totnes in southwestern Saskatchewan, Canada. Creighton is a small catchment (11·4 km^2), within which silty clay and clay loams are the two dominant soils (Gray *et al*., 1985). Approximately 85% of the basin area was cultivated land (stubble and fallow fields), and the rest of basin consisted of grassland for the periods of simulation (Gray *et al*., 1985). The basin is characterized with level open land with poor drainage and highland with rolling topography; it is drained by a grassland 'coulee' (sharp incised valley in the upland plain) from which flows the Creighton Tributary. This stream flows intermittently with most flow during and immediately after the snowmelt period. The basin has a semi-arid climate with about 300 mm of annual precipitation (Gray and Granger, 1986). The 15-year (1971–1986) average temperature and precipitation (rainfall and snowfall) during winter period (1 October–1 May) for the basin are −6·3 °C and 106 mm, respectively (Environment Canada, 2006b). Frozen soils and wind redistribution of snow develop over the winter, and snowmelt and meltwater runoff occur in the early spring. Major snowfall usually starts in November going on until April,and several episodes of snowmelt and runoff begin in March (Pomeroy *et al*., 1998). The main streamflow event of the year in this basin and elsewhere in southern Saskatchewan is normally due to the snowmelt freshet but extreme streamflows in small catchments are also due to convective rainfall events in the summer.

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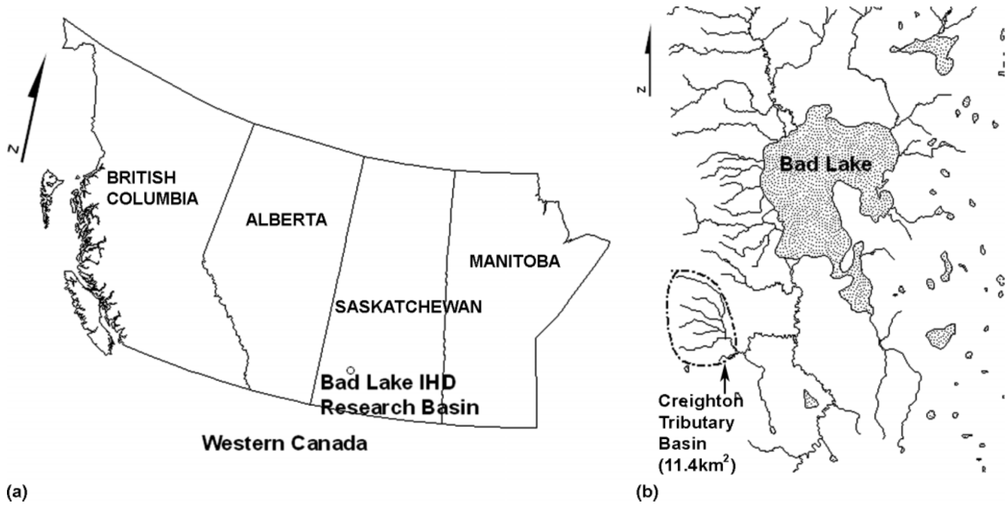

Figure 1. Location of study area: (a) Bad Lake IHD Research Basin, Saskatchewan, Canada (b) Creighton Tributary of Bad Lake Research Basin

Field observations at Bad Lake were carried out from the 1960s to the 1980s, including measurements from a Meteorological Service of Canada standard meteorological station with an on-site observer, streamflow gauges and extensive snow surveys. Specialized measurements of air temperature, relative humidity, wind speed, soil temperature, precipitation, snow depth, and radiation (direct shortwave, diffuse, and net) were obtained from the meteorological station and other sites. Snow surveys were conducted in various land uses within the basin and provided measurements of snow depth and density during the winter period and melt rate during the snowmelt period. Streamflow discharge on Creighton Tributary was monitored with a stage recorder at a weir and there were frequent velocity measurements so that reliable stage-discharge relationships could be developed for the melt period.

METHODS

The physically based CRHM was used to estimate the water balance during winter and spring period for the basin. Archived field observations were used to test the performance of the model. A sensitivity analysis to drought was conducted by adjusting snowfall and air temperature from actual winter meteorological data and changing module parameters such as initial soil saturation and vegetation height that would derive from summer meteorology and cultivation practices.

Cold Regions Hydrological Modelling Platform

CRHM is based on a modular, object-oriented structure in which component modules represent basin descriptions, observations, or physically based algorithms for calculating hydrological processes. A full description of CRHM is provided in this special issue (Pomeroy *et al.*, 2007). CRHM permits the assembly of a purposely built model from a library of processes, and interfaces the model to the basin based on a user selected spatial resolution. Hydrological processes such as wind redistribution of snow, snowmelt, snowmelt infiltration into frozen soils, and evaporation are common in prairie winter and early spring. These processes all influence spring snowmelt runoff. Snow accumulation (often called SWE) controls the amount of available snow for melting and is affected by wind in open prairie environments. Blowing snow in open environments can erode and sublimate or redistribute as much as 75% of annual snowfall from open, exposed fallow fields in southern Saskatchewan (Pomeroy and Gray, 1995). Redistribution is from open, well exposed surfaces to sheltered vegetated surfaces. The amount of surface snowmelt runoff is governed by both snowmelt infiltration and SWE. Snowmelt infiltration reduces the direct surface runoff, decreasing amount of peak flows (Norum *et al.*, 1976).

Relevant modules chosen using CRHM for these simulations included the Prairie Blowing Snow Model (Pomeroy and Li, 2000), the Energy–Budget Snowmelt Model (Gray and Landine, 1988), Gray's expression for snowmelt infiltration (Gray *et al.*, 1985), Granger's evaporation expression for estimating actual evaporation from unsaturated surface (Granger and Gray, 1989; Granger and Pomeroy, 1997), a soil moisture balance model for calculating soil moisture balance and drainage (Leavesley *et al.*, 1983), and Clark's lag and route runoff timing estimation procedure (Clark, 1945). These modules were assembled along with modules for radiation estimation and albedo changes (Garnier and Ohmura, 1970; Granger and Gray, 1990; Gray and Landine, 1987) into CRHM. This enabled the estimation of SWE after wind redistribution, snowmelt rate, cumulative snowmelt, cumulative snowmelt infiltration into INF, and actual evaporation. Actual evaporation is that calculated using the method of Granger and Pomeroy (1997) which is entirely an atmospheric energy balance and feedback approach, the approach is then modified by CRHM in that actual evaporation (E) is limited by a surface mass balance; when interception storage and soil moisture reserves are depleted evaporation cannot proceed. Snowmelt runoff

Table I. Characteristics of major module parameters for three HRUs (fallow field, stubble field, and grassland coulee) used in CRHM as the 'normal' simulation for 1974–1975 simulations and for model testing in both 1974–1975 and 1981–1982

HRU name	Area (km^2)	Soil type	Fall soil moisture (volumetric ratio)	Porosity (ratio)	Vegetation height (m)	Blowing snow fetch distance (m)	Routing lag (h)	Routing storage (day)
Fallow field	3·58	Clay Loam	0·23	0·5	0·05	1500	8	1
Stubble field	6·13	Clay Loam	0·27	0·5	0·2	2000	8	1
Grassland	1·68	Clay Loam	0·22	0·5	0·25	2000	3	0·5

(R) over the event was estimated based on a simplified conservation equation:

$$R = SWE - INF - E \qquad (2)$$

where all terms are in mm of water equivalent.

Calculations in CRHM are made on HRUs. On the basis of the major land uses in the basin and on physiography, three HRUs [fallow field, stubble field, and grassland (coulee)] were chosen for the snowmelt runoff simulation (Table I). The total snowmelt runoff from these HRUs provided the cumulative basin snowmelt runoff as:

$$R_{\text{basin}} = R_{\text{fallow}} \frac{Area_{\text{fallow}}}{Area_{\text{basin}}} + R_{\text{stubble}} \frac{Area_{\text{stubble}}}{Area_{\text{basin}}} + R_{\text{grassland}} \frac{Area_{\text{grassland}}}{Area_{\text{basin}}} \qquad (3)$$

where R_{basin}, R_{fallow}, R_{stubble}, and $R_{\text{grassland}}$ are basin snowmelt runoff, snowmelt runoff over fallow field, stubble field, and grassland, respectively; $Area_{\text{basin}}$, $Area_{\text{fallow}}$, $Area_{\text{stubble}}$, and $Area_{\text{grassland}}$ are area of basin, fallow field, stubble field, and grassland, respectively. The definition of several HRUs within a basin permits consideration of effects due to variable contributing area—HRUs are only part of the contributing area for streamflow when they produce infiltration excess or surface runoff.

CRHM test

Without calibration of parameters from streamflow observations, the simulated pre-melt snow accumulation and cumulative streamflow discharge during snowmelt from CRHM were tested against field observations of pre-melt snow depth and density in Bad Lake Research Basin during the winter of 1981–1982 and against cumulative streamflow discharge in the snowmelt period in the spring of 1975. The winter seasons of 1974–1975 and 1981–1982 were both slightly colder and wetter than the 15-year (1971–1986) average. For the snow accumulation simulations (1981–1982) only fetch length and vegetation height were important to the simulation and these were recorded from field observations made at the time. For the basin discharge simulation (1974–1975), parameters observed and reported by Gray *et al.* (1985) or noted in field observations at the time were used and are listed in Table I. Porosity was determined from soil type and values recommended by Dingman (1994). Blowing snow fetch distance was determined from maps and contemporary aerial photographs of the area. Routing lag and storage values were decided based on the HRU size and shape and landform type. The fetch length and vegetation heights in 1981–1982 were similar to that in 1974–1975.

To evaluate the performance of CRHM in simulating snow accumulation over the winter and streamflow discharge, two statistical measures, the Nash–Sutcliffe coefficient (NS) (Nash and Sutcliffe, 1970) and model bias (MB) were calculated as,

$$\text{NS} = 1 - \frac{\sum (X_o - X_s)^2}{\sum (X_o - \overline{X_o})^2} \qquad (4)$$

$$\text{MB} = \frac{\sum X_s}{\sum X_o} - 1 \qquad (5)$$

where X_o, X_s, and $\overline{X_o}$ are the observed, simulated, and mean of the observed values, respectively. A Nash–Sutcliffe coefficient equal to 1, implies that model perfectly predicts pre-melt snow accumulation and streamflow discharge with respect to observations. A value equal to 0 indicates that estimated values are indifferent to those observed. Hence, any positive value of this coefficient shows that the model has some predictive power, and better model performance is associated with higher values (Evans *et al.*, 2003). The value of model bias evaluates the ability of the model to reproduce the water balance or bias of the model, a positive bias indicates overprediction and a negative bias indicates underprediction.

Drought sensitivity analysis

A sensitivity analysis to drought was conducted to identify the influence of individual components of meteorological, soil and vegetation condition changes during drought on snowmelt runoff and governing processes. Observations at the Bad Lake IHD meteorological station from the winter of 1974–1975 were used to drive the model run (Figure 2). The winter of 1974–1975 was near 'average' with seasonal maximum snow depth and SWE of 29·9 cm and 71 mm, respectively (Gray *et al.*, 1985); the mean temperature and total precipitation of the winter of 1974–1975 were −7 °C and 129 mm, respectively. To create drought scenarios, precipitation and air temperature from observed meteorological data were adjusted from observed values over the period: 1 Oct 1974–1

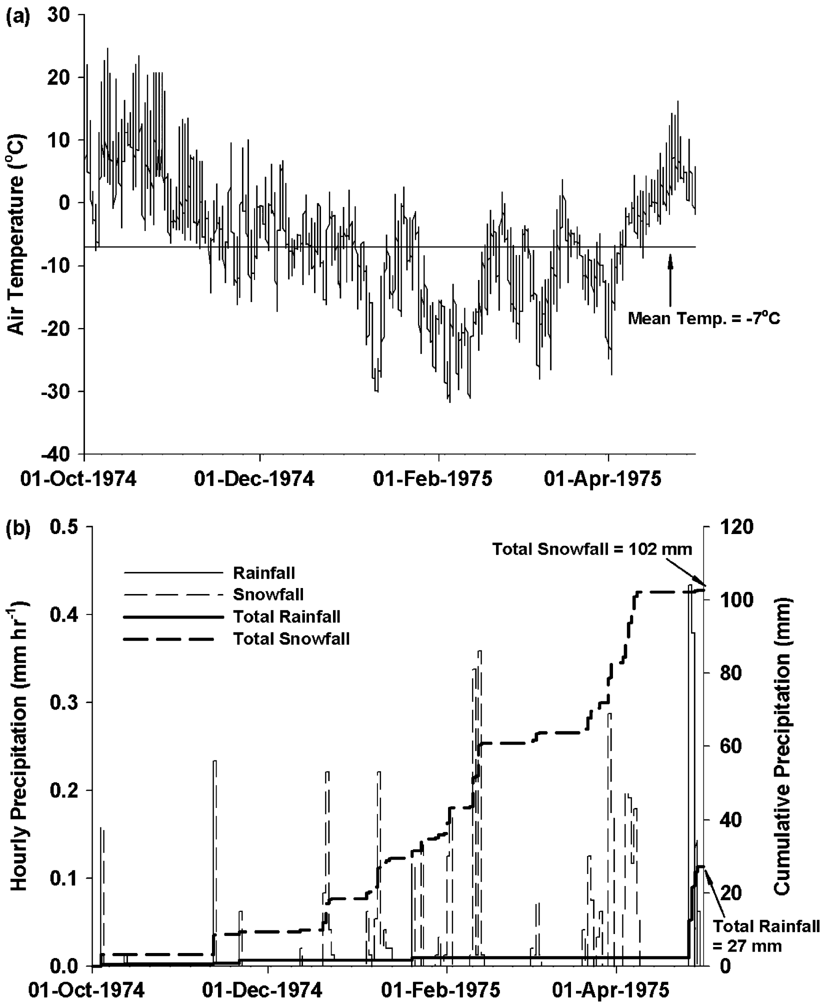

Figure 2. Meteorological observations (unadjusted) from Bad Lake Research Basin during the winter of 1974–1975: (a) air temperature, (b) precipitation as rainfall or snowfall

Table II. Drought scenarios in CRHM simulation run

Model run #	Description	Model run #	Description
1	15% decrease in precipitation	6	5 °C rise in air temperature
2	30% decrease in precipitation	7	25% decrease in volumetric soil moisture
3	50% decrease in precipitation	8	50% decrease in volumetric soil moisture
4	1 °C rise in air temperature	9	50% decrease in vegetation height
5	2·5 °C rise in air temperature	10	90% decrease in vegetation height

May 1975. Initial soil moisture content and vegetation height were changed as well. Ten drought scenarios were generated by altering one observation variable or module parameter at a time whilst holding other observation variables and module parameters constant in CRHM (Table II).

Prairie hydrological drought progression scenario

To examine the effect of combinations of major changes in meteorology, soils, and landscape on hydrological processes and snowmelt runoff during a drought, a simplified, synthetic 'prairie hydrological drought progression' was established from the general sequence of changes in precipitation, temperature, soil moisture and vegetation that was noted during the recent drought of 1999–2004, and in previous prairie droughts (Environment Canada, 2006a; Maybank *et al.*, 1995; Nkemdirim and Weber, 1999; Wheaton *et al.*, 2005). It was also informed by winter progressions of temperature and precipitation observed at nearby Rosetown during the recent droughts of 1986–1989 and 1999–2004 that show consistently lower winter precipitation and sometimes higher air temperatures than the long-term average (Figure 3). The synthetic scenario was used to adjust CRHM parameters (soil moisture, vegetation height) or variable inputs (precipitation, temperature) in a sequence

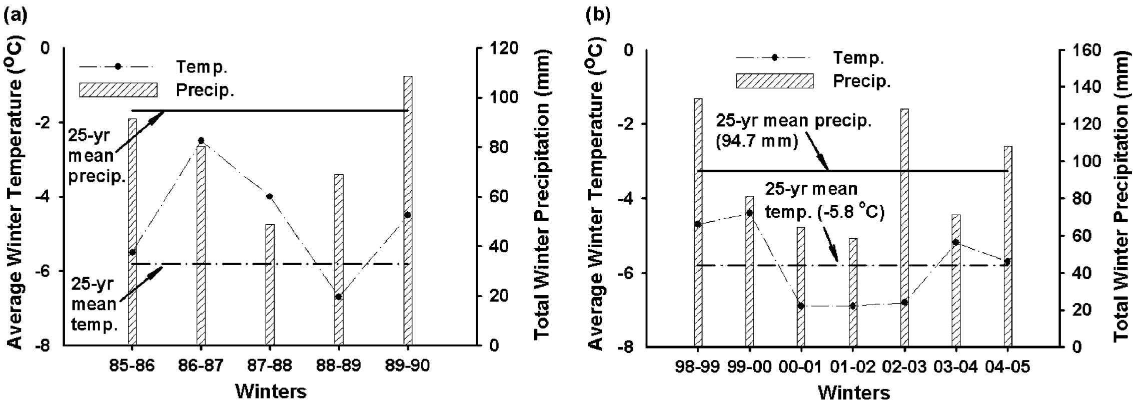

Figure 3. Average winter temperature and total winter precipitation (rainfall and snowfall) observed in Rosetown, Sask. during winters of: (a) 1985–1990, (b) 1998–2005 along with 25-year (1981–2005) winter temperature and precipitation normals

Table III. Parameters and changes in input variables for a hypothetical 'prairie drought progression'

Drought sequence	Fall volumetric soil moisture	Vegetation height (m)	Winter precipitation	Winter temperature (°C)
Winter 1 'normal'	normal	Normal	normal	normal
Winter 2 'severe'	normal	Normal	−15%	+2·5
Winter 3 'severe'	−45%	−60%	−30%	+2·5
Winter 4 'recovery'	−45%	−60%	normal	normal
Winter 5 'recovery'	normal	−35%	normal	normal

that started with no drought (Winter 1) then a winter meteorological drought (Winter 2), then the effects of summer drought on winter drought (Winter 3), then the recovery from winter meteorological drought but with antecedent conditions still in drought (Winter 4), then recovery of summer crop growing conditions fall soil moisture and winter meteorology, but not native vegetation heights (Winter 5). The progression is summarized as follows and is fully quantified in Table III:

Before a drought. **Winter 1, Normal:** No drought, fall conditions for typical for non-drought years at Bad Lake were used (early 1970s) and the 1974–1975 meteorological sequence was used.

Going into a drought. **Winter 2, Severe Winter Drought:** fall soil moisture is normal and vegetation is tall, but precipitation is lower and temperature is higher over the winter. This is the first winter of meteorological drought.

Winter 3, Severe Multi-seasonal Drought: fall soil moisture is low, vegetation is short (poor crops the previous summer reduced stubble and 'trash' on fields and native vegetation is becoming sparser), precipitation is lower, and temperatures are higher over the winter. This is the peak of the meteorological drought which has been continuous over the previous summer and fall and extends into winter.

Coming out of a drought. **Winter 4, Winter Recovery:** fall soil moisture is low and vegetation is short, but precipitation and temperature have recovered to normal winter values. The meteorological drought is 'broken' with a return to non-drought winter meteorology; however, antecedent conditions remain affected by the drought.

Winter 5, Multi-seasonal Recovery: fall soil moisture is high, precipitation is normal and temperature is normal over the winter but natural vegetation remains low. The meteorological drought ended the previous year and now soil moisture and cropped vegetation heights have also recovered, however, native vegetation heights remain short and sparse.

RESULTS

Test of CRHM for prairie snowmelt runoff

The results of 28 snow depth and density surveys on fallow and stubble fields were compared to simulations using CRHM for January–April 1982 and are shown in Figure 4. They show that the difference in snow accumulation between fallow and stubble field is well simulated over the accumulation and pre-melt period. Higher accumulation develops on the stubble because of erosion and redistribution of snow from the fallow field as blowing snow. The correct simulation of both the source (fallow) and sink (stubble) areas implies that both transport and sublimation of blowing snow are correctly estimated (Pomeroy *et al.*, 1998). In the spring of 1975, the simulated streamflow discharge (all of Creighton Tributary) due to snowmelt runoff is earlier

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than that observed by about 2 days, and somewhat greater discharge is predicted by the simulation compared to the measurement, however, the cumulative discharge curves are very similar and the timing of rapid discharge is correct (Figure 4).

To quantify differences between model and simulation, both Nash–Sutcliffe coefficient and Model Bias were calculated for fallow and stubble pre-melt snow accumulation, and cumulative streamflow discharge during snowmelt (Table IV). The Nash–Sutcliffe coefficients for snow accumulation of fallow and stubble fields, and cumulative streamflow discharge were 0·60, 0·75, and 0·90, respectively. This indicates that CRHM performs fairly well in predicting the timing of snow accumulation and very well in predicting the timing of streamflow discharge due to snowmelt (Figure 4). Fairly small values of model bias, 0·18 for both fallow snow accumulation and streamflow discharge, and 0·09 for stubble snow accumulation, represent an 18% overestimation of fallow snow accumulation and streamflow discharge and 9% overestimation of stubble snow accumulation. This implies a reasonable ability of CRHM to estimate snow accumulation in windblown prairie fields and streamflow discharge due to snowmelt over frozen soils without calibration of parameters. No calibration or adjustment of parameters was attempted from the results of these comparisons but it is presumed that the simulated hydrograph could be made to more closely mimic the observed hydrograph shape with calibration of the lag and storage parameters of the Clark unit hydrograph module.

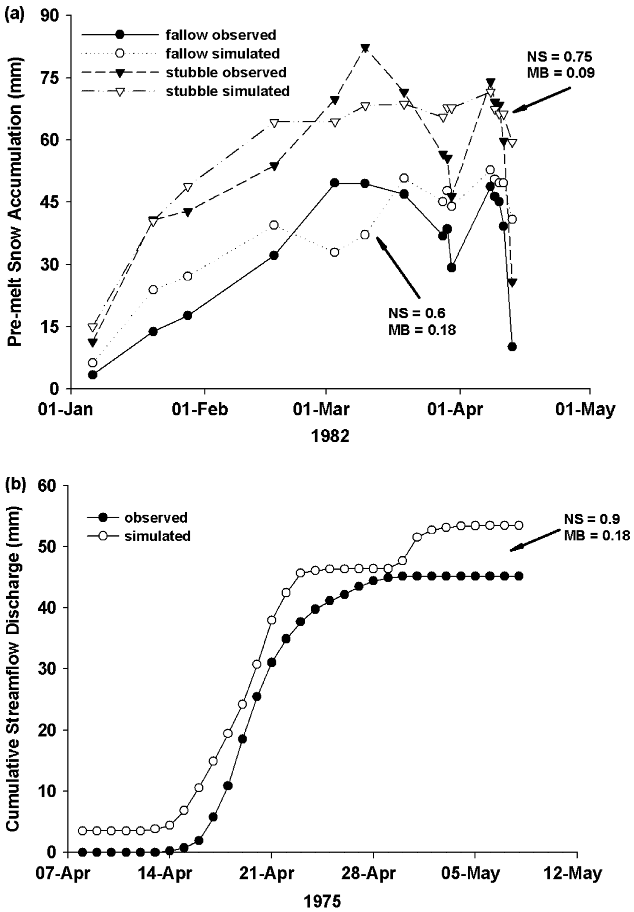

Figure 4. CRHM test: (a) simulated pre-melt snow accumulation and observed pre-melt snow accumulation in fallow and stubble fields of Bad Lake Research Basin during the winter of 1981–1982, (b) simulated cumulative streamflow discharge during snowmelt and observed streamflow discharge for Creighton Tributary, spring 1975

Sensitivity of winter hydrology to drought parameters

Corresponding to each of the ten drought scenarios (lower precipitation, higher air temperature, lower soil moisture content, and shorter vegetation), snow accumulation after wind redistribution, snow-covered period, sublimation from blowing snow, evaporation, and infiltration were estimated from CRHM for each HRU. Figure 5 shows the influence of changing the parameters and meteorological variables in these various scenarios on snow accumulation after wind redistribution (pre-melt snow accumulation). Snow accumulation is primarily sensitive to precipitation amount and less so to air temperature and vegetation height. There is no effect from changing soil moisture. When vegetation height decreased, accumulation on the fallow and especially the stubble HRU decreased linearly but accumulation on the grass HRU increased slightly because of increased redistribution of snow. So, over the basin, the effects of shorter vegetation height were somewhat compensating but overall there was decreased accumulation. When air temperature increased there was no effect on fallow snow accumulation, an increase and then a decrease (after $>2.5\,^{\circ}C$ temperature increase) in stubble snow accumulation and a decrease in grassland snow accumulation. Increasing temperatures worked to increase the frequency of rainfall at the expense of snowfall and to suppress blowing snow redistribution through increased frequency of icy and wet conditions. The suppression of blowing snow from the fallow HRU compensated for the lower snowfall amounts, for the stubble suppression of blowing snow was initially the most important and increased accumulation, but as temperature warmed further reduction in snowfall reduced accumulation. At the grassland HRU, suppression of blowing snow and reduced snowfall both worked to reduce accumulation sharply with increasing temperature. Reduced precipitation caused a moderate linear decline in accumulation on the fallow HRU but more strongly reduced accumulation at the stubble and the grassland HRU where inputs of blowing snow were also reduced with declining precipitation. The drift formation in the grassland HRU was completely suppressed once winter precipitation dropped by more than 30%.

Table IV. Nash–Sutcliffe coefficient and model bias for the CRHM test

	Nash–Sutcliffe coefficient	Model bias
Fallow Pre-melt SWE 1981	0·6	0·18
Stubble Pre-melt SWE 1981	0·75	0·09
Cumulative basin snowmelt runoff 1975	0·9	0·18

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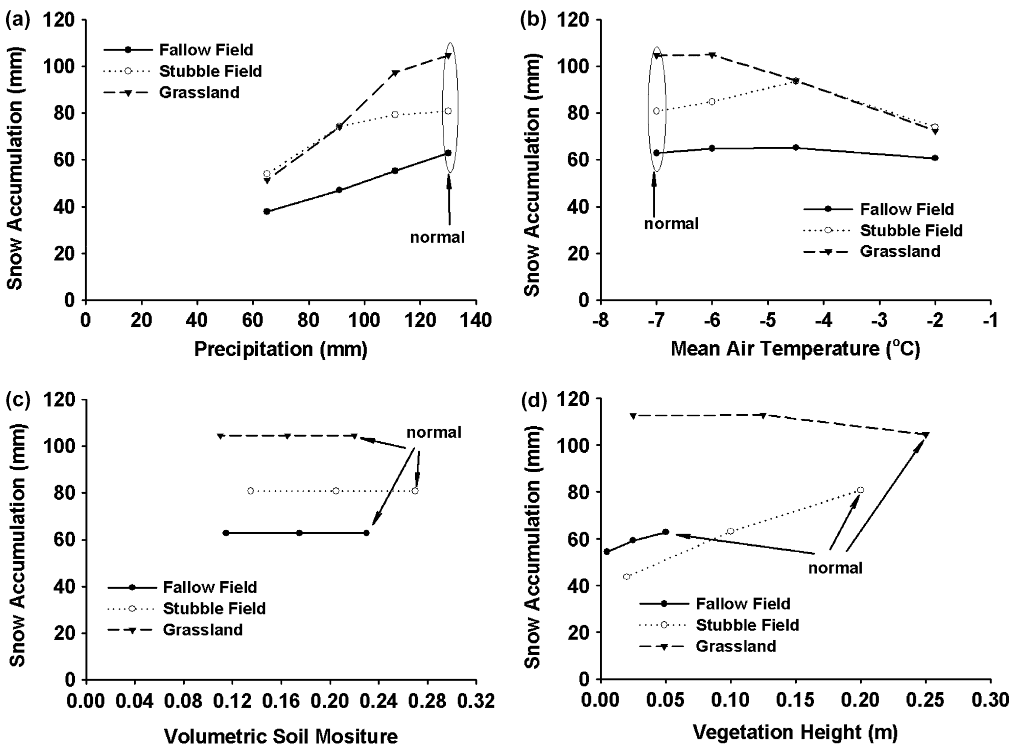

Figure 5. Drought sensitivity of snow accumulation over fallow field, stubble field and grassland HRU to changes in (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

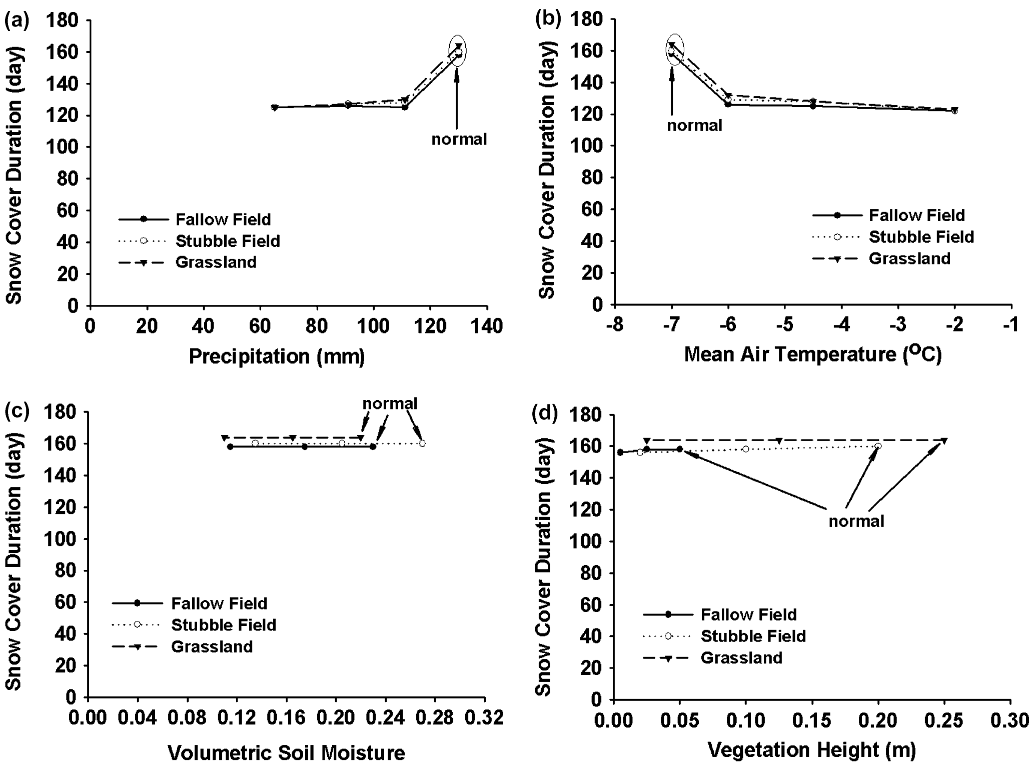

Figure 6. Drought sensitivity of snow cover duration over fallow field, stubble field and grassland HRU to changes in (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

The snow cover duration showed a much smaller sensitivity to air temperature and winter precipitation and no sensitivity to soil moisture or vegetation height (Figure 6). The initial decrease in precipitation (15%) or increase in air temperature (1 °C) caused a sharp decline in snow cover duration from 165 to 125 days, but no further reduction with changes in precipitation or air temperature was found. This is due to the 'shoulder seasons' of early and late winter being reduced in duration, but the heavier snowfalls in cold mid-winter were less

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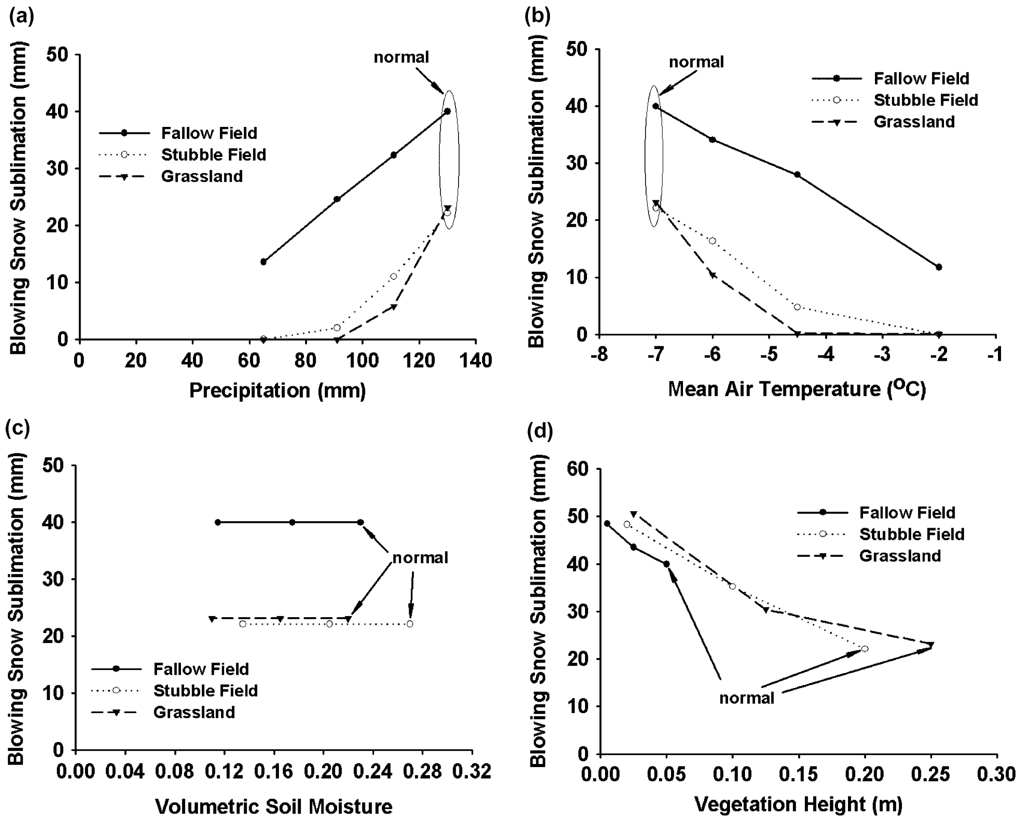

Figure 7. Drought sensitivity of blowing snow sublimation from fallow field, stubble field and grassland HRU to changes in (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

unaffected and so the core duration of snow cover was insensitive to the drier and warmer simulated conditions.

Blowing snow sublimation depends on both occurrence of blowing snow and undersaturated conditions in the atmosphere and so was very sensitive to changes in air temperature, winter precipitation and vegetation height, but not to soil moisture (Figure 7). Blowing snow was completely suppressed on the stubble and grassland HRU once the precipitation had dropped by 30% and the fallow HRU sublimation dropped sharply and linearly with precipitation. Increases in air temperature had a strong role in suppressing blowing snow sublimation with all sublimation suppressed on the grassland HRU with a 2·5 °C rise and all suppressed on the stubble HRU with a 5 °C rise. This shows that the effect of decreasing frequency of blowing snow with increasing temperatures overwhelms the effect of increasing undersaturation at higher temperatures on sublimation. Fallow HRU sublimation dropped sharply and linearly with increasing temperature. Decreasing vegetation height worked to increase sublimation, primarily for the stubble and grassland HRU where sublimation doubled with the 90% decline in vegetation. The fallow HRU sublimation changed little because vegetation there was already short and sparse.

Actual evaporation (Granger and Pomeroy's method limited by mass balance) was sensitive to air temperature, winter precipitation and sometimes to vegetation height, but not to fall soil moisture (Figure 8). An increase in air temperature of 5 °C caused a 52–172% increase in winter evaporation with the largest increase over the grassland HRU and least over the fallow HRU; suppression of blowing snow sublimation, and a longer snow free season with higher air temperatures both contributed to this response. There was initially a slight increase in evaporation as precipitation declined, presumably due to the shorter snow cover duration, but then for 50% decline in precipitation the lack of available water overwhelmed the longer period for bare ground evaporation and so evaporation declined. Declining vegetation height reduced evaporation both with respect to lower snow accumulation and subsequent water availability and to lower aerodynamic roughness height and subsequent interception losses of rainfall and wet snow. The insensitivity of evaporation to initial (fall) soil moisture indicates that there were no initial moisture supply limitations to evaporation given the adequate precipitation in the 'normal' simulation.

Infiltration was sensitive to changes in air temperature, precipitation amount, soil moisture content and vegetation height (Figure 9). The sensitivity was complex because infiltration is composed of both infiltration to frozen soils from melting snow and infiltration from rainfall into unfrozen soils and the mechanism of infiltration sometimes changed. For instance, both stubble and grassland HRU infiltration showed a substantial linear decrease with declining precipitation, with a 60–70% decrease for a 50% decrease in precipitation, but the fallow HRU infiltration showed no overall trend with precipitation. Fallow HRU infiltration increased three-fold as air temperature increased by 5 °C whereas there was little response from

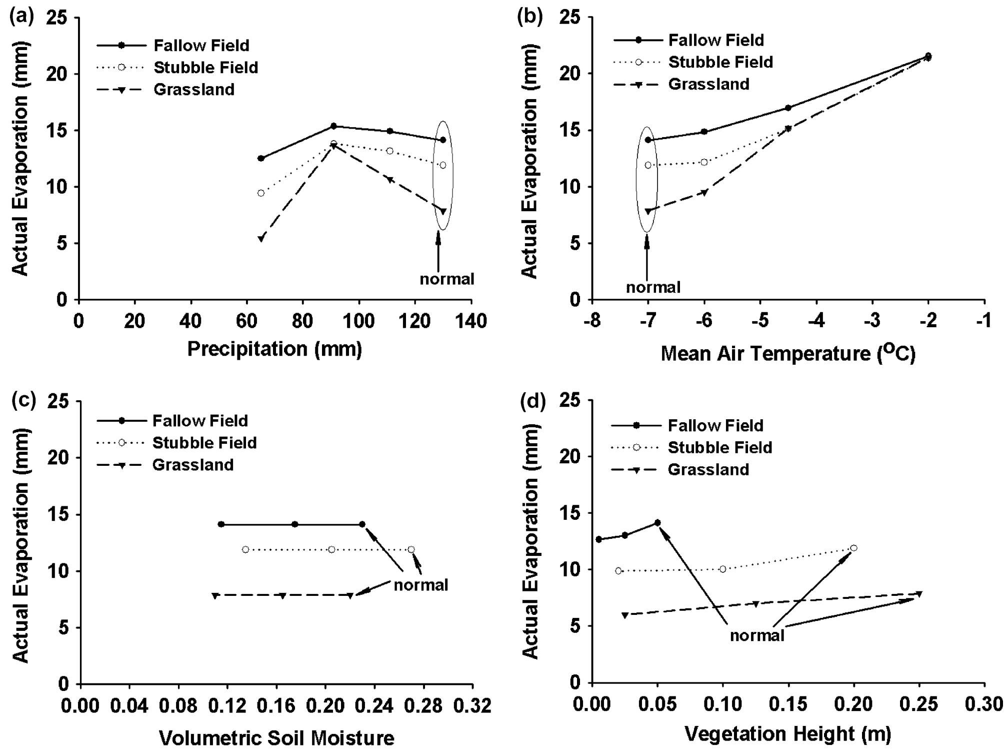

Figure 8. Drought sensitivity of actual evaporation from fallow field, stubble field and grassland HRU to changes in (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

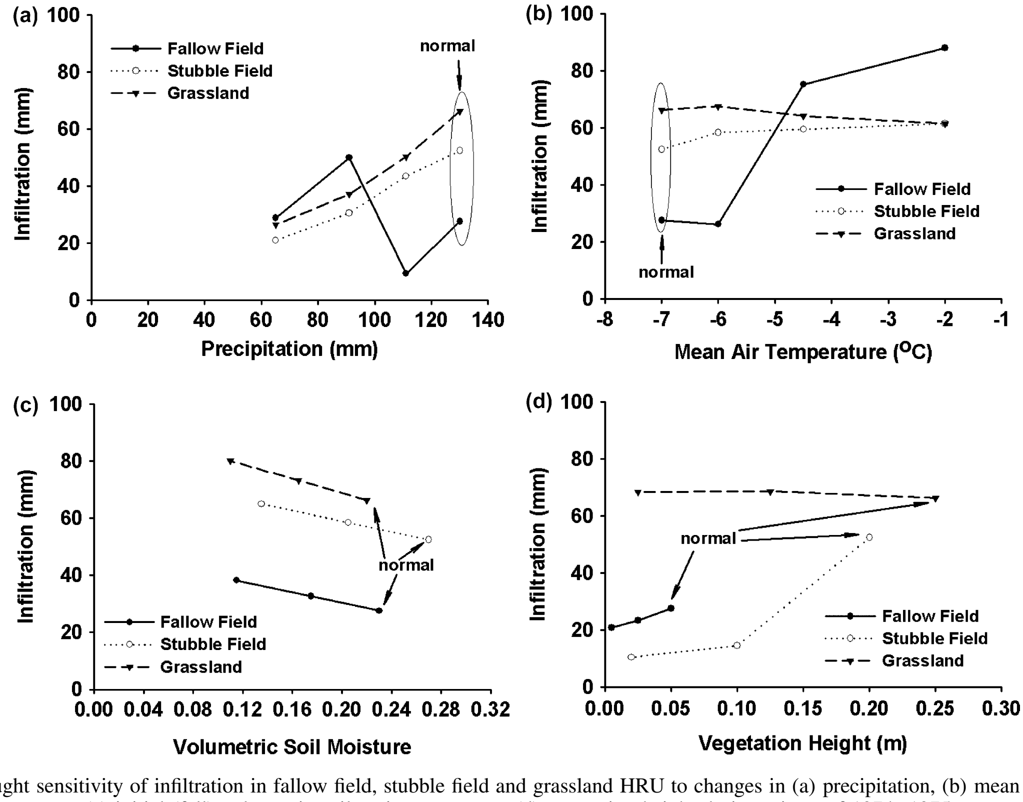

Figure 9. Drought sensitivity of infiltration in fallow field, stubble field and grassland HRU to changes in (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

the stubble and grassland HRU to changing temperature. Infiltration increased with decreasing initial (fall) soil moisture content according to Equation (1) given that porosities and snow accumulation were fixed. There was small or no response of infiltration to decreases in vegetation height for fallow and grassland HRU but a very strong decline of 75% in stubble HRU infiltration as vegetation declined by 50%.

Changes in overwinter soil moisture storage are shown in Figure 10 and display a very complex response to

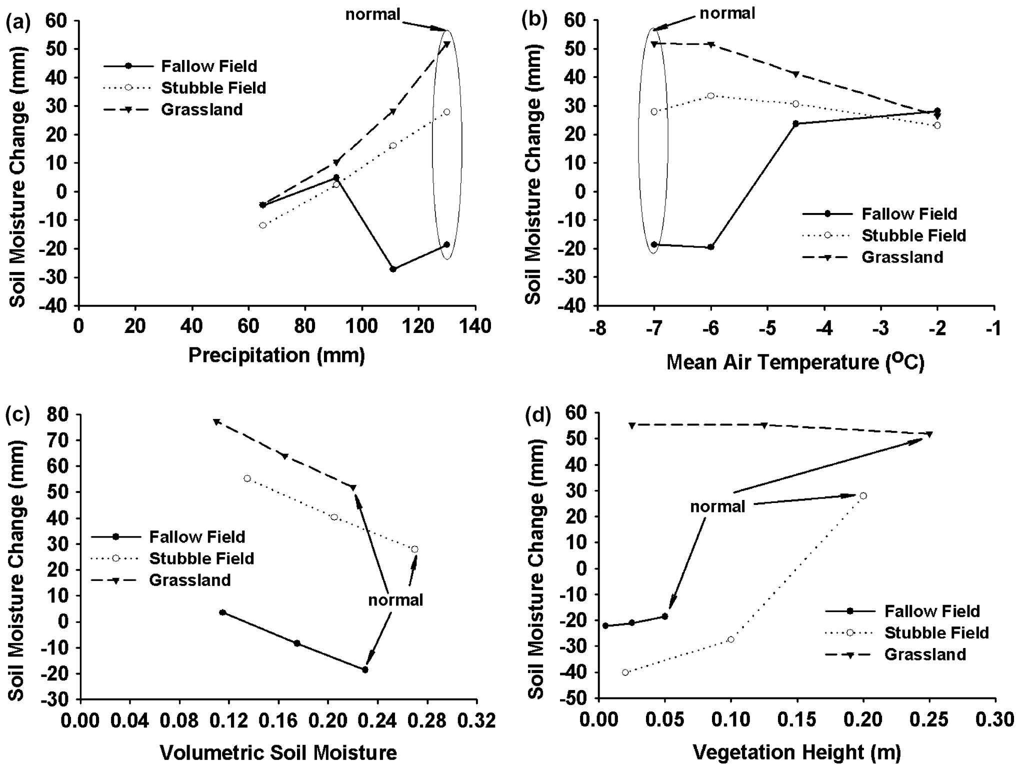

Figure 10. Drought sensitivity of overwinter soil moisture storage change (− denotes decrease) in fallow field, stubble field and grassland HRU to changes in (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

drought conditions. Under normal conditions there are large soil moisture recharges in the grassland and stubble and moderate loss from the fallow HRUs. However with a 50% decline in precipitation there are small storage losses (−5 to −15 mm) from all HRUs. With an increased air temperature there are moderate storage gains (20–30 mm) for all HRUs. Soil moisture change increases linearly as initial volumetric soil moisture declines, such that a 50% decrease in initial soil moisture results in all HRUs gaining slightly (fallow) or largely (grassland). The response to vegetation height change is most apparent for the stubble HRU where decline in vegetation height of 50% causes soil moisture to move from a moderate gain to a moderate loss—this demonstrates and confirms the benefits of snow management by using standing grain stubble in also preserving soil moisture for spring crop growth (Pomeroy *et al.*, 1990; Pomeroy and Gray, 1995).

Snowmelt runoff from the different HRUs within the basin was calculated according to Equation (2). The results (Figure 11) show snowmelt runoff over the fallow field, stubble field, and grassland HRU decreased dramatically as precipitation declined and air temperature rose and decreased with decreasing soil moisture content. Snowmelt runoff decreased linearly to zero when precipitation decreased 50% to 65 mm and when air temperature increased 5 °C to 2 °C. The decline in runoff with declining precipitation was due to winter precipitation being infiltrated and evaporated rather than forming infiltration excess and hence runoff as precipitation amounts declined in the scenarios. The strong decline in runoff with increasing air temperature was due to the conversion of winter snowfall to rainfall, the suppression of a winter snowpack, mid-winter melting of the snowpack, greater infiltration into unfrozen soils and greater evaporation losses over the winter due to the shorter snow season. In contrast, snowmelt runoff declined linearly with soil moisture as described by Equations (1) and (2). Cumulative streamflow discharge from the Creighton Tributary of Bad Lake was estimated according to Equation (3). The response of cumulative streamflow discharge to changes in snowfall, air temperature, initial soil saturation, and vegetation height was found to be similar to that of snowmelt runoff over individual HRU (Table V). Cumulative streamflow decreased 49% for a 30% decline in winter precipitation and ceased completely when precipitation decreased by 50%. Similarly, as air temperature rose by 2·5 °C, cumulative discharge dropped by 41% and ceased completely for a 5 °C temperature increase. The response to soil moisture change was more muted in that cumulative discharge decreased 23% for a 50% drop in fall soil moisture content. The streamflow response to vegetation height changes was within 1–2% of original discharge.

Basin-wide response to changes in winter precipitation, air temperature, fall soil moisture content, and vegetation height are shown in Figure 12. Pre-melt snow accumulation was more sensitive to changes in precipitation than to air temperature and vegetation height, and showed no sensitivity to soil moisture content. Winter evaporation

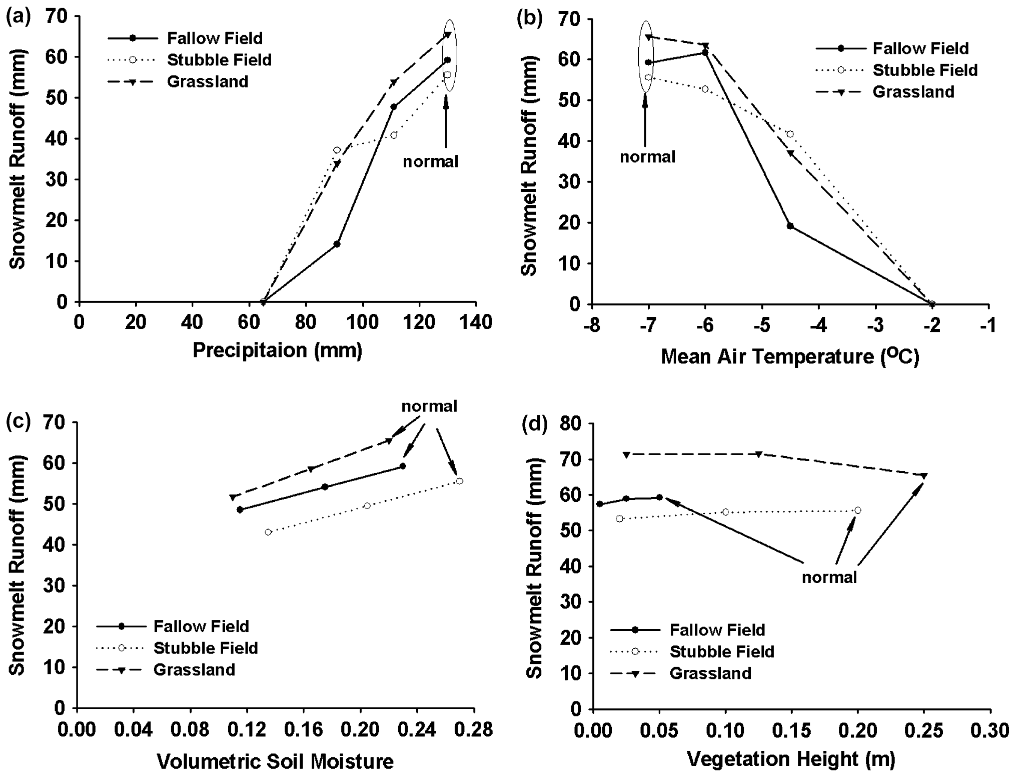

Figure 11. Drought sensitivity of snowmelt runoff over fallow field, stubble field and grassland HRU to changes in: (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

Table V. Drought sensitivity of cumulative streamflow discharge from Creighton Tributary of Bad Lake Research Basin in spring of 1975

Changes in parameter or variable	Streamflow discharge (mm)	Decrease in streamflow discharge (%)*
15% decrease in precipitation	42	21
30% decrease in precipitation	27	49
50% decrease in precipitation	0	100
1 °C rise in air temperature	51	4
2·5 °C rise in air temperature	32	41
5 °C rise in air temperature	0	100
25% decrease in volumetric soil moisture	48	11
50% decrease in volumetric soil moisture	41	23
50% decrease in vegetation height	54	−1
90% decrease in vegetation height	52	2

* Negative sign indicates increase in snowmelt runoff.

was insensitive to all parameters. Infiltration increased with increasing air temperature, precipitation and vegetation height and decreased with fall soil moisture content. Basin discharge from snowmelt was very sensitive to precipitation and air temperature and less sensitive to fall soil moisture content and vegetation height. Either a 50% drop in winter precipitation or a 5 °C rise in air temperature caused complete cessation of spring basin discharge.

Prairie hydrological drought progression

The results of seasonal simulations for the hypothetical prairie hydrological drought progression outlined in Table III are shown in Figure 13. The figure is instructive in that it shows how typical combinations of meteorological, soil and landscape factors might vary together during drought to affect the hydrological cycle. Winter 1 is the 'normal' winter with no drought, whilst Winters 2 and 3 are progressively stronger 'severe drought' and Winters 4 and 5 have progressively recovering state parameters ('recovery'). During severe drought blowing snow sublimation was largely suppressed, but it was enhanced during recovery when snowfall returned but vegetation heights remained low. As a result, even though the meteorological input of snowfall returned to normal after the severe drought period, snow accumulation did not rebound until the very end of the recovery period when vegetation (stubble, some grass) had grown to its former height and density. Interestingly, rainfall declined only slightly during the severe drought as rising air temperatures increased the proportion of precipitation falling as rain and this compensated for the overall decrease in precipitation amount. Evaporation increased somewhat during severe drought and slightly during the first year of recovery (Winter 4) but was not strongly affected by the drought progression. Infiltration was more strongly affected by drought and declined progressively through severe drought and started bouncing back during the recovery period. Basin streamflow discharge showed the greatest response to the drought progression, ceasing completely during the severe drought and not fully

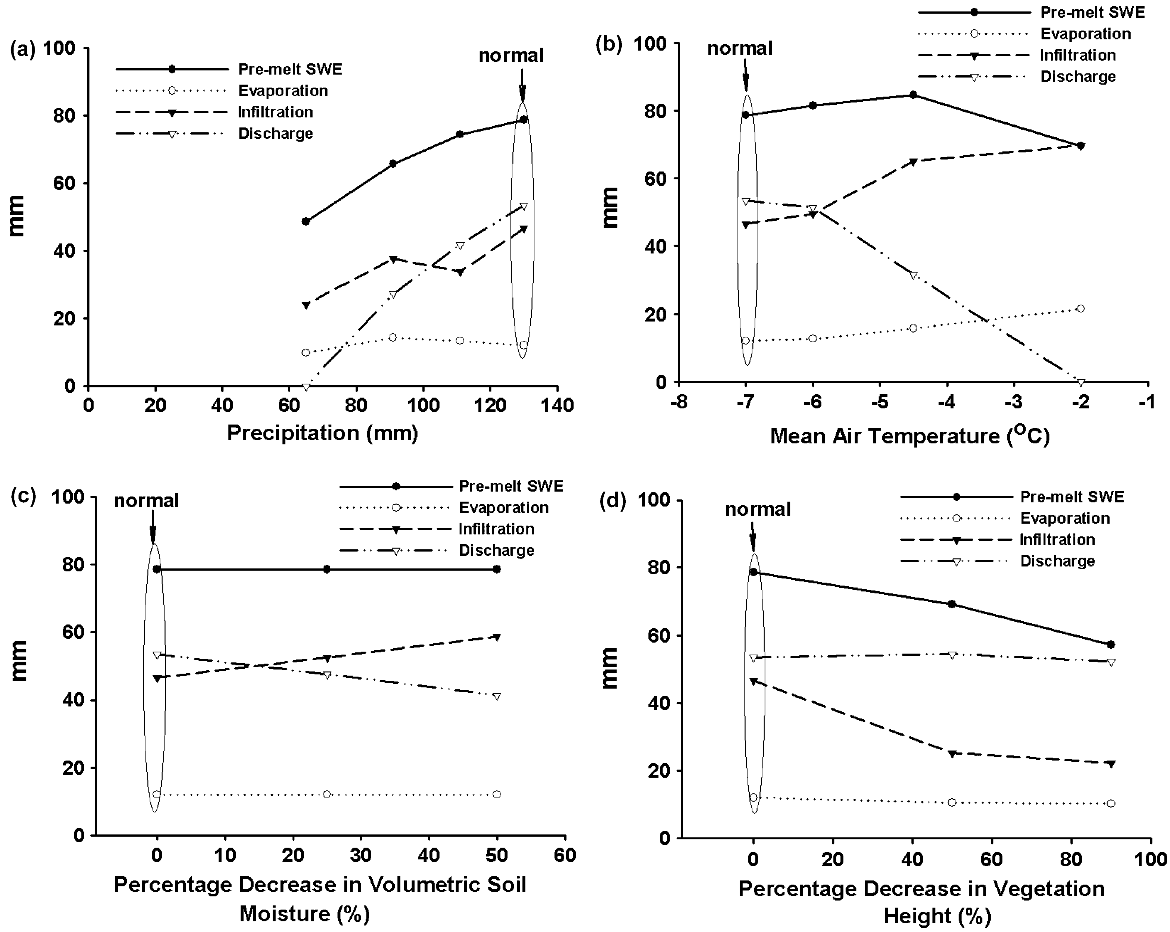

Figure 12. Drought sensitivity of basin-wide snowmelt runoff-related processes to changes in: (a) precipitation, (b) mean air temperature, (c) initial (fall) volumetric soil moisture content, (d) vegetation height during winter of 1974–1975

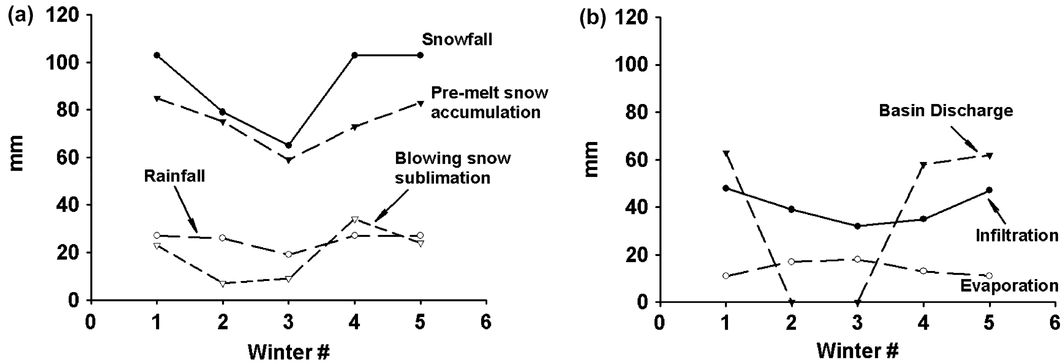

Figure 13. Prairie winter hydrological drought progression (Table III) from normal (winter 1) to severe drought (winter 2, 3) to recovery (winter 4, 5), showing the drought impact on (a) snowfall, rainfall, pre-melt snow accumulation, blowing snow sublimation and (b) infiltration, evaporation, streamflow discharge (October 1–May 1)

recovering until the end of the recovery period. The results show that even though the winter meteorological drought only occupied Winters 2 and 3, the memory in the system due to the effect of summer drought on soil moisture storage and the growth cycle of crops and native vegetation required two more winters for the hydrological drought to end. This system is very simple and has no surface storage (lakes, ponds); the presence of water bodies with large storage capacities would add even more years to the hydrological drought recovery period.

DISCUSSION

Bad Lake was not operated during a severe drought and so it is not possible to test CRHM there in drought conditions. The performance of CRHM when compared to diagnostic observations in non-drought years is encouraging but does not guarantee its good performance in severe drought scenarios. Any confidence in its performance under drought should be due to the strong physical basis of the component modules that are key to the water balances shown here and to the evaluation of the 'system' that has been assembled. It should be noted that the model includes all of the primary processes responsible for spring runoff generation in a prairie environment and has no gross errors in calculating the water balance or interaction between HRU when compared to basin observations. It should be noted that Bad Lake is only representative of the semi-arid to sub-humid region of the Canadian prairie.

There are several interesting and sometimes non-intuitive results from the sensitivity and scenario studies. One is that snow accumulation is not particularly sensitive to individual meteorological or vegetation components of a drought. This resiliency is due to blowing snow redistribution processes being very sensitive to drought meteorology and the suppression of blowing snow by meteorological conditions in a drought leaves more snowfall on the ground. Winter evaporation does not increase strongly with drought conditions and any increase in evaporation is more than compensated for by the decrease in snow sublimation. There is no strong trend in either infiltration or soil moisture change with drought because so many processes that control these fluxes counteract each other for certain specific drought conditions. Snowmelt runoff, however, decreases dramatically with drought meteorology and this is exacerbated by decreased soil moisture during drought. That snow accumulation is not strongly affected by drought but that snowmelt runoff can easily cease under typical drought meteorology and soil conditions has important implications because it means that the prairie winter hydrological system has amplified the impact of drought in translating it to streamflow generation in springtime. This has severe implications for the sustenance of prairie wetlands, ponds, lakes and the recharge of groundwater systems from these areas during drought—these implications have been described and confirmed by van der Kamp *et al.* (2003) and the mechanisms inferred by Hayashi *et al.* (2003). However, the results here provide a detailed mechanistic description of the processes involved and their sensitivity. The nature of prairie winter drought clearly needs more study—examination of observed departures from average conditions at Rosetown (Figure 3) showed some association between decreased winter precipitation and decreased winter temperature in the midst of drought—perhaps because the synoptic patterns resulting in low precipitation also caused low temperatures in winter. This association is counter-intuitive, very different from summer drought where higher temperatures are associated with lower precipitation, and needs further study to determine how general this characterization is for the Canadian prairies.

Another interesting aspect of this study is that it can show the possible changes of winter-related hydrological processes, especially the snowmelt runoff over the Canadian prairie region for the middle of this century due to anticipated climate change. Töyrä *et al.* (2005) from a review of several general circulation model simulations for the mid-21st century suggested that a warmer and 'wetter' climate is most likely for the middle and latter part of this century. The median of three most reliable scenarios (ECHAM4, HadCM3 and NCAR-PCM) suggest a rise in annual winter temperature and precipitation from the 1961–1990 average of 2·6 °C and 11·0% by 2050, and to 4·7 °C and 15·5% by 2080. The results of this scenario change in winter meteorology on spring runoff from Bad Lake Research Basin are a 24% rise and a 37% drop in years 2050 and 2080, respectively, compared to the basin runoff (54 mm) in spring of 1975. Again, the suggestion of increased prairie spring runoff under moderate climate warming shows that increased winter precipitation is more important than increased winter temperatures in spring runoff generation processes. This model result counters commonly held assumptions that climate warming must lead to drier conditions and the implicit assumption that temperature increases would overwhelm increases in precipitation in their effect on hydrology under a warming climate. Further study of the climate change impact on prairie runoff using physically based models such as CRHM is clearly needed.

The prairie drought progression scenario provides insight into the sequential development of drought and the hydrological impact when many meteorological variables and surface parameters vary in a consistent temporal pattern. As seen in examining individual factors, the early suppression and then later magnification of blowing snow sublimation loss serves to dampen the variability of snow accumulation during the course of a drought. Infiltration is slightly decreased during a drought. The main effect of the synthetic drought is complete cessation of snowmelt runoff and streamflow discharge due to the combination of lower precipitation, warmer winter temperatures and drier soils. It should be noted that this combination does not always develop during a Canadian prairie drought. The behaviour of streamflow discharge as a drought developing in this scenario is very non-linear, because all meteorological and surface factors work to diminish runoff generation at once. This inherent instability of Canadian prairie hydrology is a well known feature of the region and a reason why there is increasing reliance on water from mountain fed rivers for municipal and agricultural purposes, and is partly why the recent drought had such a severe impact on water bodies and streams in parts of the Canadian prairies.

CONCLUSIONS

It is possible to successfully simulate prairie snowmelt runoff and its driving processes using the CRHM. There is some confidence that this model can simulate drought impacts on prairie hydrology water balances because of its physical basis and lack of need for parameter calibration.

A model sensitivity study using CRHM showed that snow accumulation was relatively insensitive to decreases in precipitation and increases in air temperature during drought. This was partly due to the extreme sensitivity shown by blowing snow sublimation loss and the suppression of blowing snow under drought meteorology, which partly compensated for the reduced snowfall. However reduction of stubble vegetation height during multi-year droughts with poor harvests would work to strongly reduce snow accumulation in subsequent winters. Snow cover duration initially dropped when drought meteorology was induced but showed little sensitivity to the onset of severe drought conditions. Soil moisture storage and snowmelt infiltration changed in various directions with increasingly severe drought meteorology but generally increased as fall soil moisture declined. Evaporation increased slightly with the imposition of drought meteorology but any increased losses were more than compensated for by reduced blowing snow sublimation losses. Snowmelt runoff showed a heightened sensitivity to drought meteorology and ceased when precipitation dropped by 50% or air temperatures rose by 5 °C. The decline in stream discharge in spring resulting from this was magnified in that for every one percentage drop in precipitation, streamflow dropped 1·6%, and for every°C increase in mean air temperature, streamflow decreased 20%. In contrast, the effect of soil moisture changes on stream discharge were dampened, in that for every one percentage decrease in fall soil moisture content, streamflow declined by only ~0·5%. Vegetation height changes had no effect on stream discharge.

A prairie drought hydrology progression scenario was proposed in which severe winter drought meteorology ensued for the first two years of drought but fall soil moisture and vegetation recovered after the winter meteorology returned to normal. The combination of factors with winter precipitation declining by 15% and air temperatures rising by 2·5 °C was sufficient to cause the cessation of snowmelt runoff and streamflow discharge. The lingering hydrological impacts of the drought in the winter season subsequent to the end of the winter meteorological drought were enhanced blowing snow sublimation, and reduced infiltration in response to sparser vegetation and drier soils respectively. The system model contained little surface storage and it is likely that the addition of surface storage terms (such as prairie ponds) to the hydrological system would add a significant additional multi-year memory to the hydrological drought impacts shown here. A simple examination of the effects of climate warming anticipated by GCMs to the mid-21st century on prairie spring runoff initially showed an increase in spring runoff of 24% by 2050 followed by a decrease of 37% by 2080. This complex pattern results because the effects of increased precipitation initially overwhelmed the effects of warmer temperatures, but as warming proceeds it finally overwhelms the effects of higher winter precipitation.

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