A Review of Canadian Prairie Hydrology: Principles, Modelling and Response to Land Use and Drainage Change

Centre for Hydrology Report #2
Version 2

- By -

Xing Fang, Adam Minke, John Pomeroy, Tom Brown, Cherie Westbrook, Xulin Guo and Seifu Guangul

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1. Executive Summary

This report reviews research on the hydrological cycle, runoff generation, hydrological modelling and the influence of changes to land cover and wetlands on the same for the Canadian Prairies. The purpose of this report is to identify and examine the major processes that are responsible for prairie hydrology as well as the impacts of land cover change such as wetland drainage on water storage and on the streamflow hydrograph. The objective of this report is to propose hydrological modelling techniques; these techniques can contribute to the development of a predictive tool in the form of a prairie hydrological model. It is intent to utilize such a hydrological model to evaluate the impacts of wetland drainage and restoration as well as changes in the surrounding upland land use on downstream hydrology.

Hydrology in the Canadian Prairie region is complex and highly varied. Only one third of annual precipitation occurs over the winter and the surface snow water equivalent distribution is highly heterogeneous due to wind redistribution of snow during blowing snow storms. Blowing snow can transport and sublimate as much as 75% of annual snowfall from open prairie fields. The formation of drifts from windblown snow lengthens the spring runoff season and modulates the peak spring flows. The frozen state of mineral soils results in rapid snowmelt runoff in the springtime, which produces 80% or more of annual local runoff. The prairie region is characterized by glacially-formed depressions; these depressions fill with water to form pothole sloughs and wetlands and are very important to prairie hydrology due to their surface storage capacity. A fill-and-spill runoff mechanism is identifiable in prairie basins that are dominated by these surface depressions where flow does not commence until all storage in the depressions is filled. This results in an episodic and rapid increase in contributing area during peak runoff events. However outside of these events much of the prairie landscape is non-contributing to streamflow and even in the most extreme runoff events, some prairie basins are internally drained and never contribute to streamflow. This fill and spill phenomenon is in contrast to forms of hydrological storage found in temperate regions in which the flow rate is proportional to storage. Because of depressional storage and poorly and internally drained basins, most surface runoff in the prairie region does not contribute to the major river systems.

Hydrological processes in the prairie region are sensitive to the land cover and climate change. Wetlands can be completely dried out when surrounded by native grassland rather than agricultural fields. Droughts are frequent on the Canadian Prairies. Lower precipitation and higher air temperature are the common characteristics of droughts; surface snowmelt runoff is largely suppressed and can even completely cease when warmer (e.g. 5 ºC increase of temperature) or drier (e.g. 50% decrease of precipitation) conditions develop.

The Cold Regions Hydrological Model platform (CRHM) is a "state-of-the-art" physically-based hydrological model designed for the prairie region. 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. Preliminary tests show reasonable performance of CRHM in simulating the water balance and streamflow hydrograph for prairie regions. The model also shows capabilities to simulate impact of land use change and climate change on hydrological processes and streamflow. Further work in CHRM will be development of surface storage and surface routing models that are suitable for modelling hydrology in the prairie wetland region.

2. Introduction

The prairie region of Canada (the Prairies) lies in the southern part of provinces of Alberta, Saskatchewan, and Manitoba and is a portion of the Northern Great Plains of North America. The Prairies are characterized by relatively low precipitation especially in the southwest part due to the atmospheric flow barrier imposed by the Rocky Mountains and experience frequent water deficits and low moisture reserves (Agriculture and Agri-Food Canada, 1998). Annual precipitation in the prairie region of Saskatchewan ranges from 300-400 mm (Pomeroy *et al*., 2007a), about one third of which occurs as snowfall (Gray and Landine, 1988). The Prairies are a cold region and exhibit classical cold regions hydrology with continuous snowcover and frozen soils over much of the region in the winter. Great variation in hydrology exists across the Prairies, with fairly well-drained, semi-arid basins in the southwest part and with many wetlands and lakes development in the relatively wet north central and eastern parts.

The hydrology of the central prairie region is featured by:

- long periods of winter (usually 4-5 months) with occasional mid-winter melts (frequent in the southwest and infrequent in the northeast), with the snowcover modified by wind redistribution and sublimation of blowing snow,
- high surface runoff from the major spring snowmelt event as a result of frozen state mineral soils at the time and the relatively rapid release of water from snowpacks (Gray *et al*., 1985),
- deep soils characterized by good water-holding capacity and high unfrozen infiltration rates (Elliott and Efetha, 1999),
- most rainfall occurring in spring and early summer from large frontal systems and the most intense rainfall in summer from convective storms over small areas (Gray, 1970),
- very low levels of soil moisture, plant growth, evaporation and runoff from mid-summer to fall due to low rainfall (Granger and Gray, 1989),
- poorly-drained stream networks such that large areas are internally drained and do not contribute to the major river systems (Martin, 2001).

3. Prairie Hydrological Cycle

The main processes in the prairie hydrological cycle are shown in Figure 1. Snow is an important water resource on the Canadian Prairies. Approximately one third of annual precipitation occurs as snowfall, which produces 80% or more of annual local surface runoff (Gray and Landine, 1988). There are three scales describing the spatial variability of snow accumulation – micro (10 to 100 m), meso (100 m to 10 km), and macro (10 to 1,000 km) (Pomeroy and Gray, 1995).

In the Canadian prairie environments, snow accumulation is highly heterogeneous at micro and meso scales, due to wind redistribution of snow, also known as blowing snow. Redistribution is primarily from open, well exposed sites to sheltered or vegetated sites. There are three modes of movement involved in the transport of blowing snow – creep,

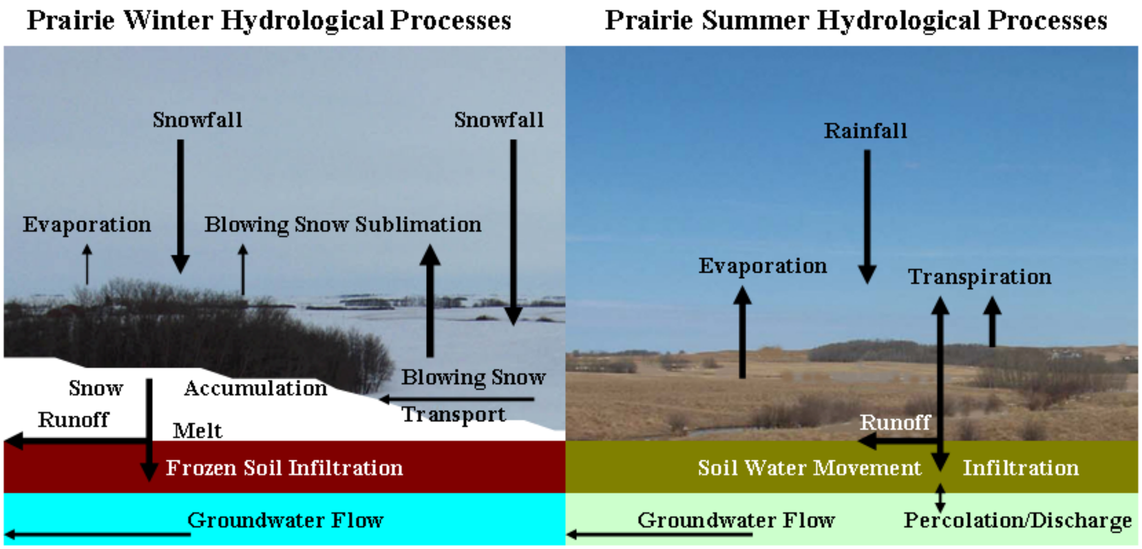

Figure 1. Prairie hydrological cycle: left – winter processes, right – summer processes.

saltation, and suspension (Pomeroy and Gray, 1995). Blowing snow transport forms snowdrifts, usually in sloughs, drainage channels or river valleys; this windblown snow provides an important source of runoff and controls streamflow peak and duration (Pomeroy *et al*., 2007a). Even though small scale heterogeneity in snow accumulation is caused by snow transport, sublimation of blowing snow contributes substantially to over-winter ablation. Seasonal sublimation of blowing snow is equivalent to 15%-40% of seasonal snowfall on the Canadian Prairies (Pomeroy and Gray, 1995). Blowing snow in the open environments can transport and sublimate or redistribute as much as 75% of annual snowfall from open, exposed fallow fields in southern Saskatchewan, how much of this can end up in a drift depends on field size, temperature, humidity and wind speed (Pomeroy and Gray, 1995).

Topography and land cover are the two major factors influencing blowing snow, because both induce variations in wind speeds which in turn affect wind redistribution of snow. In absence of vegetation cover, a leeward slope has much higher snow accumulation than does a windward slope (Steppuhn, 1981; Pomeroy and Gray, 1995). Others (e.g. Lapen and Martz, 1996) had similar findings, suggesting that the spatial distribution of snow depth in a Prairie agricultural landscape was strongly affected by the orientation of slopes and their relative position to other topographic features. Different land covers impose variations in surface roughness, which in turn causes wind speeds to change and varies distribution of snow accumulation. Pomeroy *et al*. (1990) found that southern Saskatchewan wheat stubble fields had substantially smaller losses to blowing snow than did fallow fields. Vegetation height in these agricultural fields plays an important role. As the stubble height increases from 1 to 40 cm on agricultural fields near to Regina, the loss to blowing snow decreases by 22% of the mean seasonal snowfall (Pomeroy *et al*., 1990). At Bad Lake, Saskatchewan, snow accumulation into coulees with tall shrubs was increased by approximately 50% to 100% above that contributed by the seasonal snowfall, the increase attributed to transport of blowing snow (Pomeroy *et al*., 1998).

Snowmelt is one of the most important hydrological processes on the Prairies. Melting water from snow recharges the soil moisture and groundwater storage through infiltration and replenishes reservoirs, lakes, and rivers through surface runoff (Norum *et al.*, 1976). The amount of water from snowmelt is controlled by energy exchange at the snow surface, and meltwater is produced when the snowpack is at a temperature of 0°C throughout (Male and Gray, 1981).

Seasonal rainfall occurs in the period from May to early July in the prairie region and provides water for the growth of crops. Most of the rainfall is consumed by seasonal evaporation, which leads to little surface runoff during the summer period. A primary mechanism for most rainfall events during early summer on the Prairie is the frontal weather system, while the most intense short duration rainfalls are associated with local convective storms (Gray, 1970). A detailed study of rainfall was conducted in a semi-arid area of southwestern Saskatchewan – Bad Lake Research Basin, emphasizing the spatial and temporal variability of rainfall in this region with indications for gauging network design for a prairie basin (Dyck and Gray, 1976).

Infiltration is the process by which water flows through the soils, involving a three-step sequence: entry of water into surface of soil, transmission through soil, and diminishing storage capacity in soils (Musgrave and Holtan, 1964). The process is governed by the combined influence of gravity and capillary forces (Gray, 1970; Kane and Stein, 1983). In the winter, infiltration is into frozen soils on the prairies. Through intense field studies of snowmelt infiltration carried out on agricultural land in Saskatchewan, Gray *et al.* (1985) proposed a classification that separates the frozen prairie soils to three groups depending on their infiltrability: restricted, limited, and unlimited. It is a widely used classification (e.g. Gray *et al.*, 1986, 2001; Zhao and Gray, 1997). Unlimited class soils are extremely porous and include coarse sands and gravels or cracked clays; all melting water infiltrates to these soils resulting in no surface runoff. Restricted class soils are completely saturated, and include wet heavy clays or soils with an impeding layer such as an ice lens resulting from a mid-winter melt; as a result they are impermeable so that all snowmelt water goes to runoff. Limited class soils are unsaturated soils of moderate texture that can infiltrate 10% - 90% of snowmelt water with higher quantities for drier soils. The unsaturated frozen soil system is by far the most complex soil system with two solid components: soil and ice, and two fluid components: water and air and yet it is very common in natural systems (Kane and Stein, 1983). Infiltration into such a system is a complicated process involving coupled heat and mass flow with phase changes (Zhao and Gray, 1997; Zhao *et al.*, 1997; Gray *et al.*, 2001). Infiltration into unsaturated frozen soils can be described by two regimes: a transient regime and a quasi-steady-state regime. The transient regime follows immediately after the application of water; the infiltration rate decreases rapidly during this regime. The transient regime is followed by quasi-steady-state regime in which changes in the infiltration rate with time are relatively small (Zhao and Gray, 1997; Zhao *et al.*, 1997). The soil moisture content in the previous fall and the occurrence of major melt events in mid-winter are extremely important in controlling snowmelt runoff rates in the subsequent spring (Pomeroy *et al.*, 2007a). Field investigations conducted in the western and central regions of Saskatchewan indicate that the infiltration of snowmelt water is enhanced up to six-fold by sub-soiling, or ripping, to a depth of 60 cm (Pomeroy *et al*., 1990). In the summertime, infiltration from rainfall is enhanced when the soil is thawed and this usually leads to minimal surface runoff. Limited runoff is due to both infrequent rainfall, and rainfalls of short duration as well as the high infiltration capacity of prairie soils which are most often unsaturated at the surface.

Evaporation is driven by the net radiation to the surface and by convection of water vapour from wet surfaces to the relatively dry atmosphere. In winter, both radiation and convection are relatively low, thus water loss to the evaporation during winter is much lower compared to the summer evaporation. During summer, evaporation consumes most rainfall on the prairies and occurs quickly via direct wet surface evaporation from water bodies, rainfall intercepted on plant canopies and wet soil surfaces; it occurs more slowly as unsaturated surface evaporation from bare soils and as transpiration from plant stomata (Granger and Gray, 1989). Evaporation, directly from bare soils and indirectly by transpiration, withdraws soil moisture reserves and eventually results in soil desiccation if there are no further inputs of water from rain or groundwater outflows. On average, seasonal evaporation loss is close to seasonal rainfall in Saskatchewan, with amounts less than rainfall occurring in exceptionally wet or cool years, especially in the east and north of the agricultural region. Locally higher rates of evaporation occur from sloughs and wetlands, where redistribution of spring snowmelt runoff water into topographic depressions or groundwater outflows provide for wet surface conditions through much of the summer (van der Kamp *et al*., 2003).

Groundwater recharge usually occurs at depression sites such as sloughs, wetlands and pothole lakes through the infiltration in the soil columns and deep percolation below the rooting depth (Hayashi *et al*., 2003). Much of the infiltration water for shallow groundwater recharge is exhausted by evaporative transpiration by plants (Parsons *et al*., 2004). This leads to very low and steady deep groundwater flow rates; 5-40 mm a reported range of annual groundwater recharge rates in the prairie (van der Kamp and Hayashi, 1998).

4. Prairie Runoff Generation

The Canadian Prairies are characterized by numerous small depressions such as sloughs, wetlands and dugouts. These water bodies are often internally drained resulting in closed catchment (Hayashi *et al*., 2003), and there is a lack of connection amongst them as well as to the main prairie streams. Where there is internal drainage in normal conditions these catchments are termed non-contributing areas (Godwin and Martin, 1975) and are illustrated in Figure 2. Other areas do drain to streams.

The seasonality of Prairie water supply is marked. In fall and winter, the water is stored as snow, and lake and ground ice; in early spring, the water supply is derived from rapid snowmelt resulting in most runoff; in late spring and early summer, water stored as soil moisture and surface water, sustained by rainfall. Snowmelt water contributes 80% or

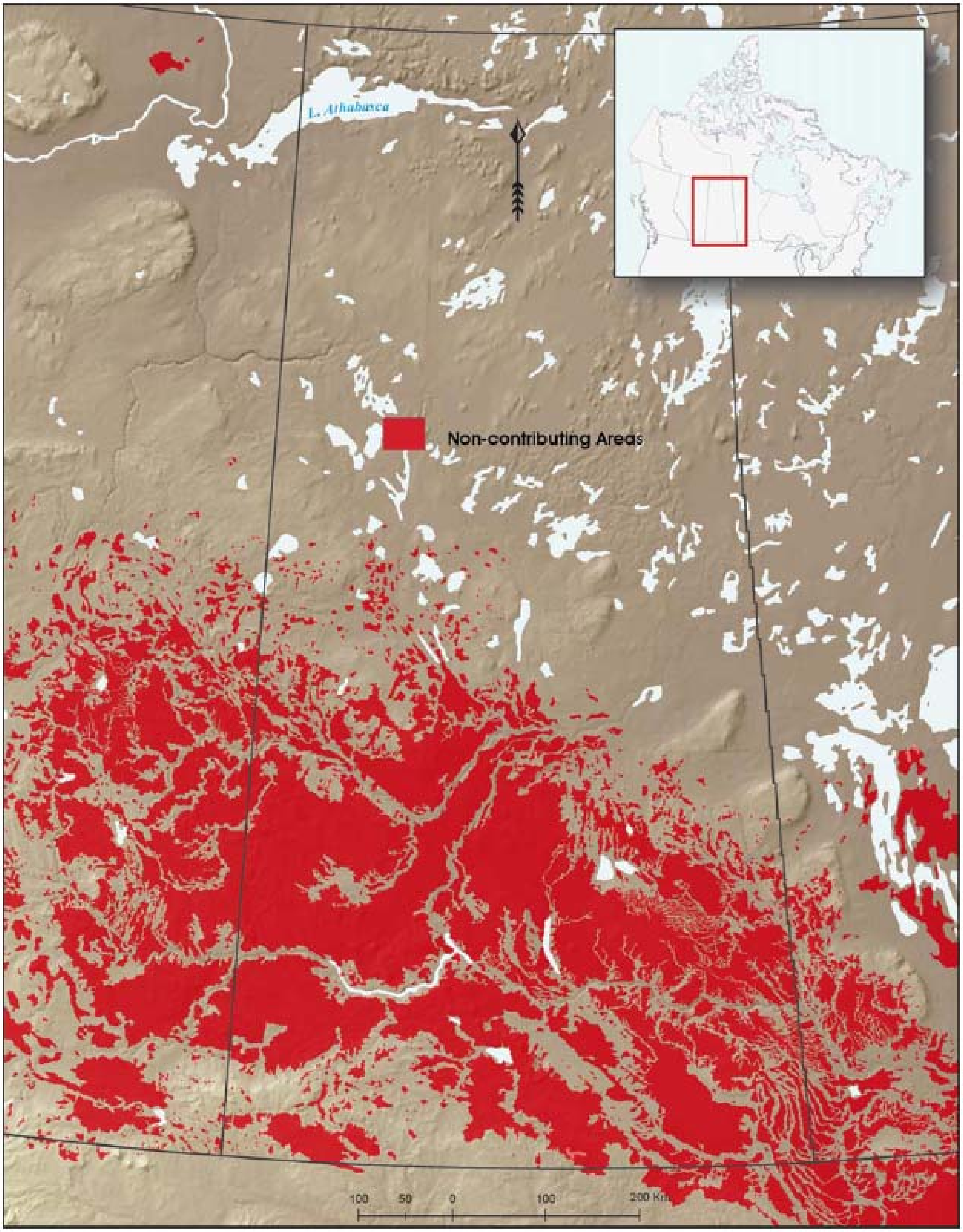

Figure 2. Non-contributing areas of drainage basins as delineated by PFRA (image from Pomeroy *et al*., 2007a).

more of annual surface runoff for Prairie streams (Gray and Landine, 1988). However, due to the aridity and gentle topography of the prairie, natural drainage systems are poorly developed, disconnected and sparse, resulting in surface runoff that is both infrequent and spatially restricted (Gray, 1970). Recent drainage activities have increase runoff to streams and wetlands in some regions. Flow in the main prairie rivers originates in the Rocky Mountains. Pomeroy *et al*. (2007a) indicated that the springtime peak stream discharge of North Saskatchewan River at Deer Creek is related to the prairie and parkland snowmelt, whereas the peak discharge in the early summer is due to snowmelt in the Rocky Mountains.

Surface runoff generation is affected by the climate variation over the Prairies. Most of the Canadian prairie region lies in the Palliser Triangle, where droughts frequently develop, and water resource is under tremendous stress during droughts. A synthetic drought analysis at a typical semi-arid prairie site suggests that spring stream discharge drops substantially under warmer and drier conditions and ceases completely when precipitation decreases by 50% or air temperature rises by 5 ºC during drought (Fang and Pomeroy, 2007). Water supply to wetlands, which are excellent wildlife habitat, is in shortage during drought due to lower discharge of surface runoff from vicinity catchments. Figure 3 shows the water level of a typical wetland pond at St. Denis NWA nearby Saskatoon and shows much lower spring pond level in drought period compared to non-drought period due to suppressed snowmelt runoff.

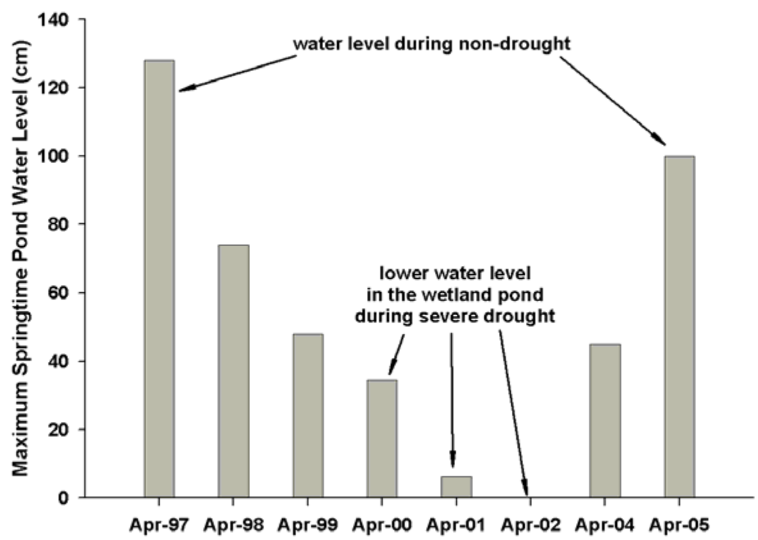

Figure 3. Observed springtime water levels in pond 109, St. Denis NWA during 1997-2005. Water level data acquired from van der Kamp *et al.*, 2006.

5. Land Cover and Wetland Effects on Prairie Hydrology

Land cover exerts great control on the prairie hydrology and is an essential factor affecting snow accumulation process in the southern agricultural region. Table 1 shows blowing snow sublimation and transport losses for fallow and stubble fields in various parts of Saskatchewan (Pomeroy and Gray, 1995). Stubble fields have substantially less loss to transport and sublimation of blowing snow when comparing to fallow fields, about

Table 1. Blowing snow transport and sublimation losses for fallow and stubble fields of 1 km length in Saskatchewan (winter is Nov – Mar).

Station	Snowfall (mm)	Winter Temp. (^{o}C)	Winter Wind Speed (m/s)	Land Use	Transport (mm)	Sublimation (mm)	Accumulation (mm)
Prince Albert	103	-11.6	4.5	Stubble	9	24	70
				Fallow	13	28	62
Yorkton	125	-10.6	4.7	Stubble	10	19	96
				Fallow	16	29	80
Regina	113	-8.9	6.0	Stubble	21	38	54
				Fallow	41	46	26
Swift Current	132	-6.7	6.6	Stubble	15	29	88
				Fallow	38	38	56

31-60% and 14-24% less transport and sublimation losses, respectively. Thus, the seasonal snow accumulation in stubble fields approximately ranges 1.1-2.1 times that in fallow fields with greater difference in the more southern agricultural region.

Figure 4 shows the land cover effect on a prairie water balance for a year with near normal precipitation. The water balance is for Creighton Tributary of the Bad Lake basin in south-western Saskatchewan and water balance for each land cover and a spatially area-weighted average for the whole basin are shown. 85 % of basin area (11.4 km^2) is cultivated field (Gray *et al.*, 1985), with 31% summer fallow (fall-spring 1974-75) then grain crop (summer 1975), 54% stubble (fall-spring 1974-75) then grain crop (summer

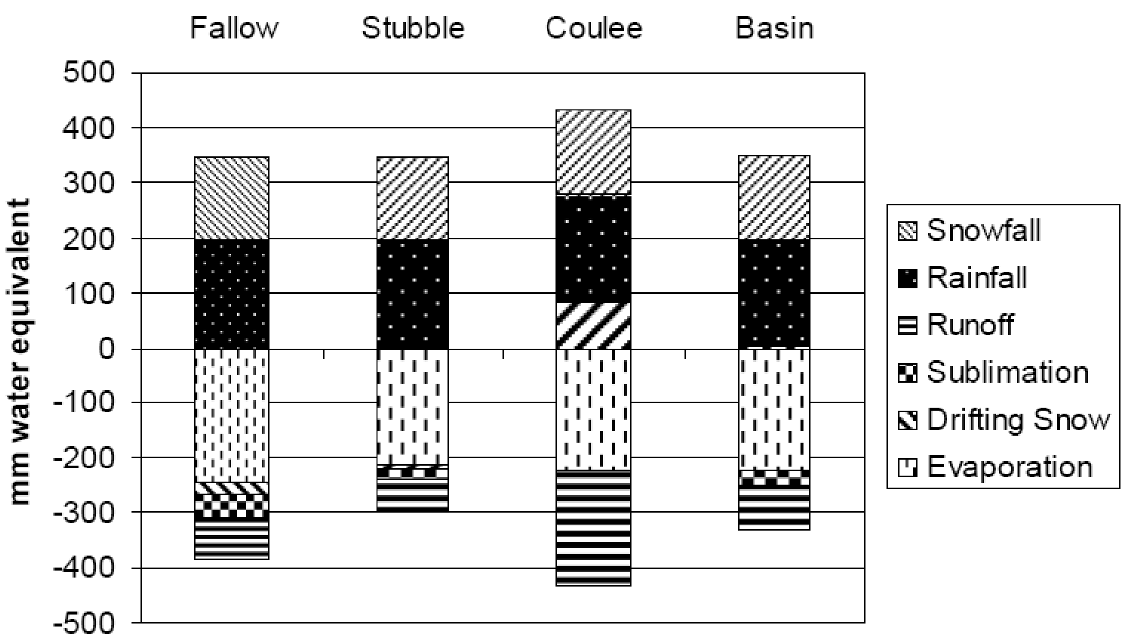

Figure 4. Water balance of Creighton Tributary of Bad Lake Basin, Saskatchewan (from Pomeroy *et al.*, 2007a).

1975), 15% brush coulee. The water balance was calculated from observations and model output from a prairie hydrological process model (Pomeroy *et al*., 2007a). Over the winter, coulee ('sink' of blowing snow) gains snow by 85 mm, while fallow fields and stubble fields ('source' of blowing snow) lose snow by 22 mm and 8 mm, respectively. During spring snowmelt, runoff is 5 times higher than infiltration on fallow fields due to nearly saturated frozen soils, and 6 times higher than infiltration in the coulee due to steep slopes and frozen soils, but infiltration is slightly higher than runoff on the stubble fields due to dry soils from the previous year's cropping, resulting in no runoff. During growing season in early June, the fallow field has lost a net 85 mm of soil moisture since fall, while soil moisture in the stubble field remains relatively constant with infiltration from snowmelt balancing evaporation in fall, spring and early summer. Runoff is dominated by the coulee, with the fallow field also making a large contribution (Pomeroy *et al*., 2007a).

Land use in the catchments of prairie wetlands plays a vital role in controlling the water supply to the wetlands. Studies have been conducted in a typical prairie wetland area – St. Denis NWA, Saskatchewan to investigate the effects of land use on the surface soil hydraulic properties and hydrological processes (Bodhinayake and Si, 2004; van der Kamp *et al.*, 2003). They indicate that wetlands within the areas converted from cultivated fields to grasslands have lower surface runoff from melting snow compared to the wetlands within the cultivated areas. Even though more snow is trapped in wetlands within the grasslands area (van der Kamp *et al.*, 2003), the soil cracks or macropores develop after long undisturbed period, resulting in unlimited soil infiltrability that enables all melting water to infiltrate into soils in the grasslands (Gray *et al.*, 2001). This leads to the drying out of wetlands (van der Kamp *et al*., 1999). This finding has application in prairie wetland hydrology. In the semi-arid south-western part of prairie, the water shortage stress would be alleviated if agricultural cropland is kept in the vicinity of wetland, so that water supply in wetland would be generated from spring snowmelt runoff. While in the relatively wet north-eastern part of prairie, the magnitude of annual spring flooding could be lessened if the wetlands are surrounded by undisturbed, natural grasslands, so that the wetland can remain relatively dry in order to store the storm water and reduce the flood peaks.

On the Canadian prairie regions, landscapes are characterized by formerly glaciated depressions. These depressions vary in size from 1 m^2 to 100 km^2 and retain water on the surface; small depressions are regarded as surface roughness terms by hydrologists (Hansen, 2000); whereas large depressions are seen as wetlands or lakes. These large depressions are very important elements in surface hydrology as they have great retention capacity (Hayashi *et al*., 2003). These large surface storage depressions have a significant influence on the basin's runoff repose and timing; the wetlands and lakes in the upper-basin can delay runoff at the basin outlet substantively (Spence, 2000). Surface runoff water flows from the basin headwater during snowmelt and heavy rainfall to the wetlands and lakes in the upstream areas. This water remains stored in the upper basin until surface storage is filled. After storage is satisfied, additional surface water spills and flows to the wetlands and lakes further downstream in a cascade fashion, ultimately reaching the outlet. This is common in the basins that are dominated by wetlands or lakes and is identified as a fill-and-spill runoff system (Spence and Woo, 2003). The fill-and-spill runoff mechanism is affected by both the location and size of available surface storage in a basin, which in turn is influenced by hydrological processes within the landscape units in the basin and inputs from upstream landscape units (Spence and Woo, 2006). These landscape units are termed hydrological response units by their behaviour and are described by the temporal pattern of their functions (Spence and Woo, 2006). The fill-and-spill runoff system can also affect the contributing area of basin. Spence (2006) found that besides the magnitude of precipitation and evaporation loss and storage capacity of lakes, contributing area of basin varied as a function of relative location of lakes within basin and size of the lakes, which resulted in different basin streamflow regimes.

6. Hydrological Modelling Techniques

In the southwest and central regions of Saskatchewan, methods and techniques to estimating various hydrological processes – snow accumulation, snowmelt, infiltration, and snowmelt runoff were investigated over several decades by the Division of Hydrology, University of Saskatchewan. Their emphasis was on developing proper modelling methods of estimating the areal snow water equivalent for a shallow prairie snow cover (Gray *et al*., 1979) and establishing techniques for predicting runoff from snowmelt (Gray and O'Neill, 1973; Erickson *et al*., 1978). Despite the relatively simple physics in these earlier modelling methods and techniques, their principle of involving in partitioning a watershed into several land units based on the landscape and land cover types forms the baseline for subsequent prairie hydrological modelling.

The Cold Regions Hydrological Model platform (CRHM) is the latest prairie hydrological model development and is the "state-of-the-art" physically-based hydrological model for the prairie region. 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. The component modules have been developed based on the results of over 40 years of research by the University of Saskatchewan and National Water Research Institute in prairie, boreal, mountain and arctic environments. A full description of CRHM is provided by Pomeroy *et al*., (2007b). 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.

Current modules in CRHM that is relevant to the prairie hydrological processes include:

- Prairie Blowing Snow Model (Pomeroy and Li, 2000),
- Energy-Budget Snowmelt Model (Gray and Landine, 1988),
- Gray's expression for snowmelt infiltration (Gray *et al*., 1985; 2001),
- Green-Ampt infiltration and redistribution expression (Ogden and Saghafian, 1997),
- 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. Additional modules will incorporate wetland/lake storage and spill and dynamic drainage network. Detailed calculating equations for some of above modules are described in the following section.

Snow Accumulation

The Prairie Blowing Snow Model (PBSM) developed by Pomeroy (1988) assembles the physically based algorithms to estimate seasonal snow accumulation on Canadian Prairies. The algorithms update snow accumulation flux, Q_A, by calculating saltation, suspension and sublimation rates of blowing snow described by Pomeroy *et al.* (1998) as:

$$Q_A(F) = P - \frac{Q_R(F) - Q_R(0)}{F} - Q_E \qquad [1]$$

where P is precipitation rate (kg/m/s), F is fetch distance of blowing snow (m), Q_R is downwind blowing snow transport (saltation and suspension) flux (kg/m/s) and Q_E is sublimation flux (kg/m/s). A control volume concept shown in Figure 5 is applied to estimate the mass fluxes of blowing snow over a certain part of landscape (Pomeroy and Li, 2000).

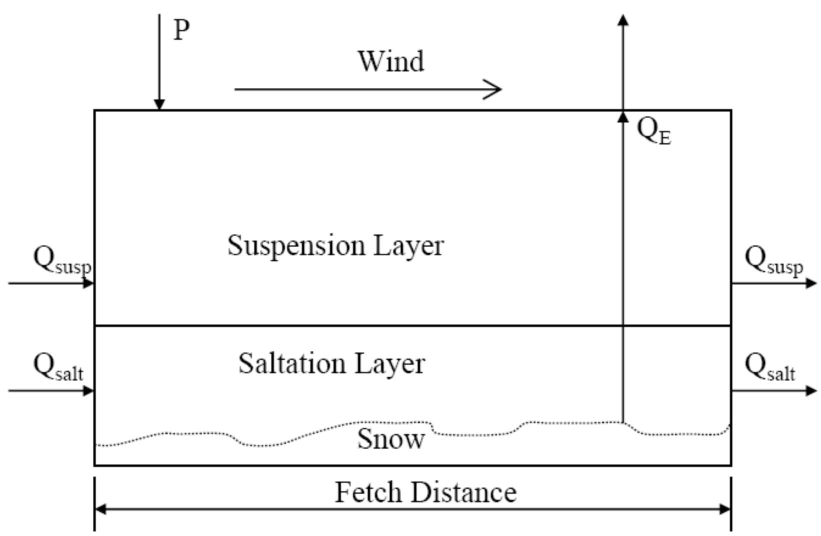

Figure 5. Cross-sectional view of control volume for blowing snow mass fluxes.

Individual fluxes of blowing snow are described by Pomeroy *et al*. (1993) in the Prairie Blowing Snow Model (PBSM), which calculates the fluxes of transport by the following equation:

$$Q_R = Q_{salt} + Q_{susp} \quad [2]$$

where Q_{salt} and Q_{susp} are the fluxes of saltation and suspension, respectively. The flux of saltation is estimated by the following equation:

$$Q_{salt} = \frac{C_{salt}\rho u_t^*}{u^* g}\left(u^{*2} - u_n^{*2} - u_t^{*2}\right) \quad [3]$$

where: Q_{salt} = saltation transport rate (kg/m/s),
C_{salt} = empirical constant (0.68 m/s),
ρ = atmospheric density (kg/m^3),
g = gravitational acceleration (m/s^2),
u^* = atmospheric friction velocity (m/s),
u_n^* = friction velocity applied to non-erodible surface elements (m/s), and
u_t^* = friction velocity applied to the snow surface (m/s).

Equation [3] was formulated by Pomeroy and Gray (1990) to apply Bagnold's framework (1954) for calculating the transport rate of saltating sand to saltating snow. Equation 3 includes the total atmospheric shear stress, τ, shear stress applied to non-erodible surface elements, τ_n, and shear stress applied to erodible surface elements, τ_t, to estimate the mean weight of saltating snow. The various types of shear stress are related to the corresponding friction velocity – u^*, u_n^*, and u_t^*. The friction velocity is calculated as a function of the wind profile:

$$u^* = \frac{u_z k}{\ln[\frac{z}{z_0}]} \quad [4]$$

where: u_z = wind speed at height of z (m/s),
k = von Kármán's constant (0.4),
z_0 = aerodynamic roughness height (m).

The non-erodible friction velocity, u_n^*, was found to equal zero for complete snow covers without exposed vegetation; the threshold friction velocity, u_t^*, is the friction velocity at which transport ceases and was found in the range of 0.07-0.25 m s^{-1} for fresh, loose snow and higher ranger of 0.25-1.0 m s^{-1} for old, dense snow (Pomeroy and Gray, 1990).

The flux of suspension is estimated by the following equation:

$$Q_{susp} = \frac{u^*}{k}\int_{h^*}^{z_b} \eta(z)\ln(\frac{z}{z_0})dz \quad [5]$$

where: Q_{susp} = suspension transport rate (kg/m/s),
u^* = atmospheric friction velocity (m/s),
k = von Kármán's constant,

z_b = upper boundary of suspension (m),
h^* = lower boundary of suspension (m),
$\eta(z)$ = mass concentration of suspended snow (kg/m^3) at height z, and
z_0 = aerodynamic roughness height (m).

Pomeroy and Gray (1990), fitting wind speed measurements, found an expression for the aerodynamic roughness height over complete snow covers as a function of friction velocity, u^*:

$$z_0 = 0.1203 \frac{u^{*2}}{2g} \qquad [6]$$

The lower boundary of suspension, h^*, which defines the saltation-suspension interface was found to relate to friction velocity, u^*:

$$h^* = 0.08436 u^{*1.27} \qquad [7]$$

Pomeroy and Male (1992) developed an expression relating the mass concentration of suspended snow to height, z, and friction velocity, u^*:

$$\eta(z) = 0.8 \exp[-1.55(4.784 u^{*-0.544} - z^{-0.544})] \qquad [8]$$

PBSM models the sublimation rate based on the energy equilibrium of radiation, convection of snow particles, water vaporation from snow particles, and sublimation (Schmidt, 1991). The sublimation rate is approximated by the following equation:

$$\frac{dm}{dt} = \frac{2\pi r\sigma - \dfrac{Q_r}{\lambda_T T Nu}\left(\dfrac{L_s M}{RT} - 1\right)}{\dfrac{L_s}{\lambda_T T Nu}\left(\dfrac{L_s M}{RT} - 1\right) + \dfrac{1}{D\rho_s Sh}} \qquad [9]$$

where: r = radius of a snow particle possessing mass m,
σ = ambient atmospheric undersaturation of water vapour with respect to ice,
Q_r = radiative energy absorbed by the particle,
L_s = latent heat of sublimation (2.838×10^6 J kg^{-1}),
M = molecular weight of water (18.01 kg mol^{-1}),
λ_T = thermal conductivity of the atmosphere ($\lambda_T = 0.00063T + 0.0673$),
Nu = Nusselt number,
R = universal gas constant (8313 J mol^{-1} K^{-1}),
T = ambient atmospheric temperature,
ρ_s = saturation density of water vapour at T,
D = diffusivity of water vapour, and
Sh = Sherwood number.

A relation between the threshold friction velocity, u_t^*, and air temperature (°C), T at 2-m height was derived by Li and Pomeroy (1997a):

$$u^*_t = 0.35 + \frac{T}{150} + \frac{T^2}{8200} \qquad [10]$$

The equation [10] provides a direct method to calculate threshold condition for blowing transport from meteorological data. Li and Pomeroy (1997b) found that the probability of blowing snow occurrence to follow a cumulative normal distribution with regard to the mean wind speed, u_{mean}, and the standard deviation δ of wind speed, u, as:

$$p = \frac{1}{\delta\sqrt{2\pi}} \int_0^u \exp^{-\frac{(u_{mean}-u)^2}{2\delta^2}} du \qquad [11]$$

They found the mean wind speed and the standard deviation of wind speed were as functions of snow conditions and air temperature based on extensive study on the Canadian prairies. For wet snow, the values of 21 and 7 m s^{-1} were found for the mean and standard deviation of wind speed, respectively. For dry snow packs, the mean and variance of wind speed were associated to air temperature (°C), T, and snow age index, I, as follows:

$$u_{mean} = 0.365T + 0.00706T^2 + 0.9I + 11.2 \qquad [12]$$

$$\delta = 0.145T + 0.00196T^2 + 4.3 \qquad [13]$$

Equation [11] allows the application of blowing snow fluxes calculation from the meteorological data and provides a technique for approximating areal blowing snow fluxes from a point. The estimation of snow mass balance using this technique was conducted in the Canadian arctic and prairie regions (Pomeroy and Li, 2000).

Snowmelt

Snowmelt involves phase changes and hence the energy equation is traditionally taken as the physical framework for snowmelt estimations (Granger *et al.*, 1977; Gray and Landine, 1988). The energy equation is based upon the law of conservation of energy to a control volume of snow, and this volume has a snow-ground interface and a snow-air interface as its lower and upper boundaries, respectively (Figure 6).

The energy budget for calculating snowmelt involves energy and mass fluxes in radiation, convection, conduction, and advection along with a change in internal energy. The equation for the energy budget is expressed as:

$$Q_m = Q_n + Q_h + Q_e + Q_g + Q_p + Q_A - \Delta U/\Delta t \qquad [14]$$

where: Q_m = energy flux available for snowmelt (W m^{-2}),

Q_n	= net radiation flux (W m^{-2}),
Q_h	= convective flux of sensible heat (W m^{-2}),
Q_e	= convective flux of latent heat (W m^{-2}),
Q_g	= conductive flux of ground flux (W m^{-2}),
Q_p	= advection from rain in vertical direction (W m^{-2}),
Q_A	= small-scale advection from patches of soils in horizontal direction (W m^{-2}),
$\Delta U/\Delta t$	= rate of change in internal energy (W m^{-2}).

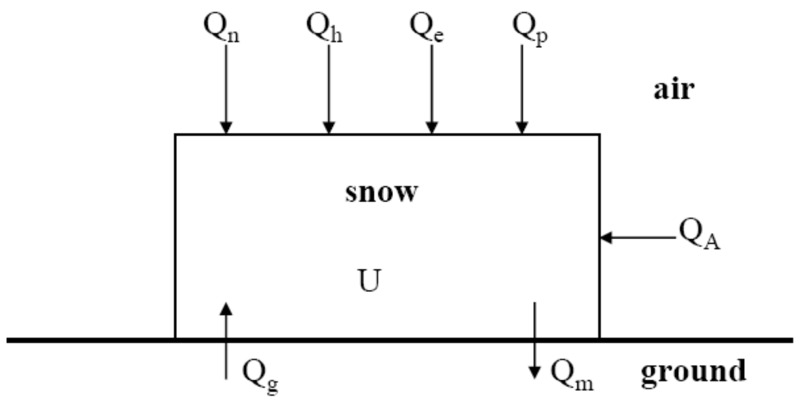

Figure 6. Cross-sectional view of control volume for snowmelt energies.

Individual terms in the energy budget equation can be determined by existed equations. Net radiation, Q_n, is the total of the net short-wave, Q_{sn}, and net long-wave, Q_{ln}, expressed as:

$$Q_n = Q_{sn} + Q_{ln} \quad [15]$$

where the net-long wave is normally negative and the energy fluxes directed towards the snow pack are considered as positive. The net-short wave is the sum of the incident short-wave flux, Q_s, received by the surface and reflected short-wave flux, Q_r, by the surface; the reflectance short-wave energy is a fraction of the incident short-wave, which is expressed as the albedo of snow, α_s, normally in the range of 0.65-0.95 depending on the age of snow. Thus, the net-short wave is expressed as:

$$Q_{sn} = Q_s(1 - \alpha_s) \quad [16]$$

The incident short-wave flux, Q_s, is total of its direct beam, Q_{drs}, and diffuse, Q_{dfs} components, expressed as:

$$Q_s = Q_{drs} + Q_{dfs} \quad [17]$$

With cloud cover, the amount of direct beam short-wave flux is reduced and found to be a function of the direct beam short-wave radiation under clear sky, Q_{dro}, as:

$$Q_{drs} = Q_{dro}[a + b(\frac{n}{N})^c] \quad [18]$$

where n/N is the sunshine ratio; a, b, and c are coefficients and are found to equal 0.024, 0.974, and 1.35, respectively, for the southwestern prairie region of Saskatchewan (Granger and Gray, 1990). The direct beam short-wave radiation under clear sky, Q_{dro}, is estimated by the expression developed by Garnier and Ohmura (1970). The atmospheric constituents (e.g. dust particles, water droplets, and ice crystals) reduce the transmissivity for beam radiation and increase the scattering and diffusion. Granger and Gray (1990) derived a relation for estimating the diffuse flux with cloud cover, Q_{dfs}, as a function of diffuse radiation under clear sky, Q_{dfo}, and the sunshine ratio, n/N, as:

$$Q_{dfs} = Q_{dfo}[2.68 + 2.2(\frac{n}{N}) - 3.85(\frac{n}{N})^2] \quad [19]$$

The clear-sky diffuse radiation, Q_{dfo}, can be the expression derived by Granger and Gray (1990) that relates Q_{dfo} to the atmospheric pressure ratio, cosine of the angle of incidence of the sun's rays on a slope, and day of year. The net-long wave flux, Q_{ln}, is the sum of the downward radiation emitted by the atmosphere, $Q_{l\downarrow}$, and the upward radiation emitted by the surface, $Q_{l\uparrow}$. Due to the influence of diurnal changing temperature on the internal energy content of shallow snowcovers, it is important to incorporate the long-wave into the snowmelt estimation. Granger and Gray (1990) developed an expression for calculating the net long-wave flux under cloud cover, Q_{ln}, for the southwestern prairie region of Saskatchewan as:

$$Q_{\ln} = Q_{\ln o}[0.25 + 0.75(\frac{n}{N})] \quad [20]$$

where n/N is the sunshine ratio; Q_{lno} is the clear sky net long-wave radiation, estimated by the following expression relating to the clear sky short-wave radiation, Q_{so}:

$$Q_{\ln o} = -4.25 - 0.24Q_{so} \quad [21]$$

Male and Gray (1981) outlined simplified bulk transfer expressions for calculating convective sensible heat flux Q_h and latent heat flux Q_e:

$$Q_h = D_h U_z (T_a - T_s) \text{ and} \quad [22]$$
$$Q_e = D_e U_z (e_s - e_a) \quad [23]$$

where:
D_h = bulk transfer coefficient for sensible heat transfer (kJ/m^3·°C),
D_e = bulk transfer coefficient for latent heat transfer (kJ/m^3·°C),
U_z = wind speed at a reference height (m/s),
T_a, T_s = temperature of the air and the snow surface, respectively (°C),
e_s, e_a = vapour pressures of the air and snow surface, respectively (mb).

Ground heat flux Q_g and internal energy U are estimated by the following equations (Male and Gray, 1981):

$$Q_g = -k(\partial T_g/\partial z) \quad [24]$$

where:
- k = thermal conductivity (W m^{-1} °C^{-1}),
- T_g = ground temperature (°C),
- Z = depth (m).

$$U = d(\rho_i C_{Pi} + \rho_l C_{Pl} + \rho_v C_{Pv}) T_m \quad [25]$$

where:
- d = depth of snow (m),
- ρ = density (Kg/m^3),
- C_P = specific heat (KJ/kg·°C),
- T_m = mean snow temperature (°C), and
- i, l, v = ice, liquid and vapor phases, respectively.

When rain falls on a melting snow pack where the rain does not freeze, the advection flux from rain is estimated by the following equation (Male and Gray, 1981):

$$Q_p = 4.2(T_r - T_s)P_r \quad [26]$$

where:
- T_r = temperature of the rain (°C),
- T_s = snow temperature (°C), and
- P_r = depth of rain (mm/day).

When patches of bare soils and snowcovers appear, the amount of energy transferred from the patchy bare soils to patchy snowcovers is an important in melting snow. This horizontal advection flux Q_A is estimated as the change of sensible heat from upwind edge of snowcovers to downwind edge of snowcovers over the length of snowcovers (Granger *et al.*, 2002). That is:

$$Q_A = aX_s^b \quad [27]$$

where:
- X_s = horizontal distance between upwind and downwind edges of snow covers,
- a, b = Weisman's dimensionless coefficients.

The amount of melt can be calculated from Q_m by the equation:

$$M = \frac{Q_m}{\rho_w B h_f} \quad [28]$$

where:
- ρ_w = density of water (1000 kg m^{-3}),
- B = fraction of ice in a unit of wet snow (0.95 → 0.97),
- h_f = latent heat of fusion of ice (333.5 kJ kg^{-1}).

Infiltration

i. Snowmelt Infiltration

On Canadian Prairies Granger *et al.* (1984) developed an empirical equation for estimating cumulative infiltration (INF) of limited infiltrability frozen soils based on the SWE and average pre-melt water and ice content of 0-300 mm soil layer (S_I). Gray *et al.* (1985) successfully derived its expression as:

$$INF = 5 \cdot (1 - S_I) \cdot SWE^{0.584} \qquad [29]$$

Zhao *et al*. (1997) developed a physically-based finite difference numerical model, HAWTS (Heat And Water Transport in frozen Soils). The model estimates moisture movement related to sensible and latent heat transfers in frozen soils based on a set of partial differential equations. Zhao and Gray (1999) developed a parametric form of the HAWTS model that describes cumulative infiltration into frozen unsaturated soils of limited infiltrability as:

$$INF = C \cdot S_0^{2.92} \cdot (1 - S_I)^{1.64} \cdot \left(\frac{273.15 - T_I}{273.15} \right)^{-0.45} \cdot t_0^{0.44} \qquad [30]$$

where *C* is a constant and is found to be 2.10 and 1.14 for the prairie soils and forest soils, respectively (Gray *et al.*, 2001). S_0 is the surface saturation (mm^3/mm^3) and is assumed to be 0.75 and 1.00, approximately equal to 1 in most of situations when infiltration rate is low and snowmelt is rapid (Gray *et al.*, 2001). S_I is the average soil saturation of top 400 mm soil layer at the start of infiltration (mm^3/mm^3) and is estimated from the average pre-melt volumetric moisture content (water + ice) (θ_I) divided by the soil porosity (Φ). θ_I can be approximated from the fall soil moisture θ_f based on empirical expressions developed on Canadian agricultural region (Gray *et al.*, 1985) as:

$$\theta_I = -5.08 + 1.05\theta_f \quad \text{(for fallow fields)} \qquad [31]$$

$$\theta_I = 0.294 + 0.957\theta_f \quad \text{(for stubble fields)} \qquad [32]$$

T_I is the average initial temperature of top 400 mm of soil (K). t_0 is the infiltration opportunity time (h) and approximately equals the time required to melt the snow cover and is estimated by the following equation:

$$t_0 \cong t = \frac{SWE}{Melt} \qquad [33]$$

The assumptions made for the parametric equation are that soil is homogeneous and isotropic, distributions of initial soil temperature and moisture are uniform, and melting water has a constant head at the soil surface (Zhao and Gray, 1999).

ii. Rainfall Infiltration

The rainfall infiltration rate according to the original Green and Ampt (1911) for a single ponding event is

$$f_p = -K_s\left(\psi_f \frac{\theta_s - \theta_i}{F} + 1\right) \qquad [34]$$

where: f_p = potential infiltration rate (cm/s),
K_s = soil saturated hydraulic conductivity (cm/s),
ψ_f = suction at wetting front (negative pressure head),
θ_i = initial moisture content (dimensionless),
θ_s = saturated moisture content (dimensionless) and
F = cumulative infiltration (cm).

Rearranging the [20] gives the cumulative infiltration *F* as a function of inilftration rate f_p

$$F = -\psi_f \frac{\theta_s - \theta_i}{1 + \dfrac{f_p}{K_s}} \qquad [35]$$

Evaporation

Actual evaporation from natural non-saturated surfaces is estimated according to the evaporation expression of Granger and Gray (1989)

$$E = \frac{\Delta G k_c (Q^* - Q_g)}{(\Delta G + \gamma)} + \frac{\gamma G E_A}{(\Delta G + \gamma)} \qquad [36]$$

where: Q^* = net radiation (W m^{-2}),
Q_g = ground heat flux (W m^{-2}),
K_c = unit conversion coefficient to provide evaporate in mm/day,
Δ = slope of the saturation vapour pressure curve (kPa/°C),
γ = psychrometric constant (kPa $^{\circ}C^{-1}$),
G = relative evaporation,
E_A = drying power (mm/day).

Details for each term above are discussed by Granger and Gray (1989; 1990).

CRHM Delineation of Basin

The modelling scheme in CRHM is shown in Appendix A and enables the estimation of snow accumulation after wind redistribution, snowmelt rate, cumulative snowmelt, cumulative snowmelt infiltration into unsaturated frozen soils, cumulative infiltration into unfrozen soils, actual evaporation, soil moisture changes, and surface runoff.

The model runs on basins which are broken down into Hydrological Response Unit (HRU) spatial units. HRU are the smallest land units having definable hydrological characteristics and need not directly drain to any stream or wetland – hence they can be used to determine contributing area for medium sized basins. Their relative physical location in the basin is important to simulations but need not be exactly known. CRHM takes in input parameters for HRU GIS files or other formats.

Preliminary Model Tests

Preliminary tests have been conducted and two statistical measures, the Nash-Sutcliffe coefficient (NS) (Nash and Sutcliffe, 1970) and Model Bias (MB) were used to evaluate the performance of CRHM in simulating snow accumulation over the winter and streamflow discharge (Figure 7 and Figure 8).

Figure 7 shows the CRHM test in a semi-arid prairie region, Bad Lake Research Basin, Saskatchewan. The NS 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. Relatively 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.

Figure 8 illustrates the CRHM test in a prairie hummocky depression region, wetland 109 at St. Denis NWA, Saskatchewan. The simulated pre-melt snow accumulation (SWE) is close to the observation on the cultivated fields (source area) and wetland (sink area) for the spring of 2000; the simulations of both source and sink areas are generally in accordance with the observations for the springs of 2001, 2003, and 2006 (Figure 8(a) and Figure 8(b)). This implies that both transport and sublimation of blowing snow were correctly simulated for springs of these years. The cumulative pre-melt SWE for the basin was estimated from the cultivated field and wetland and the statistical indicator Model Bias (MB) was also calculated to quantify the differences between observation and simulation of cumulative basin pre-melt SWE (Figure 8(c)). The values of MB are 0.04 and 0.06 for the springs of 2000 and 2003, respectively, representing an overestimation of 1.2 mm and 3.6 mm pre-melt SWE for 2000 and 2003. This suggests that CRHM performed well in predicting the cumulative pre-melt snow accumulation due to wind redistribution for these two years. Values of MB, -0.18 and -0.12 for 2001 and 2006 indicates that an underestimation of 8 mm and 12 mm pre-melt SWE for 2001 and 2006. This suggests only moderate discrepancies to observations of pre-melt SWE for these years. The cumulative snowmelt runoff to the basin of Wetland 109 was compared to the

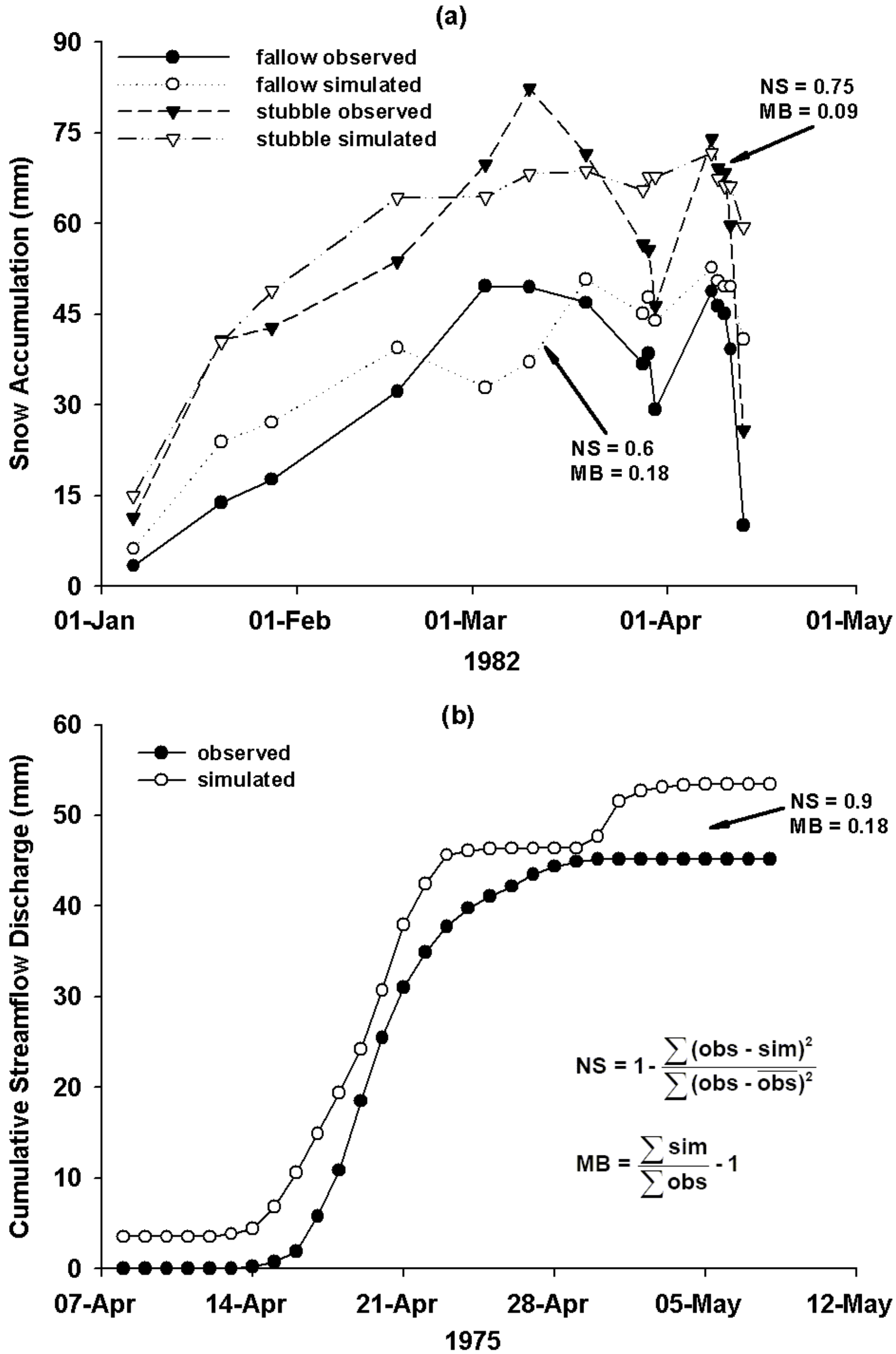

Figure 7. CRHM test: (a) simulated pre-melt snow accumulation and observed pre-melt snow accumulation in fallow and stubble fields of Bad Lake Research Basin during the winter of 1981-1982, (b) simulated cumulative streamflow discharge during snowmelt and observed streamflow discharge for Creighton Tributary, spring 1975.

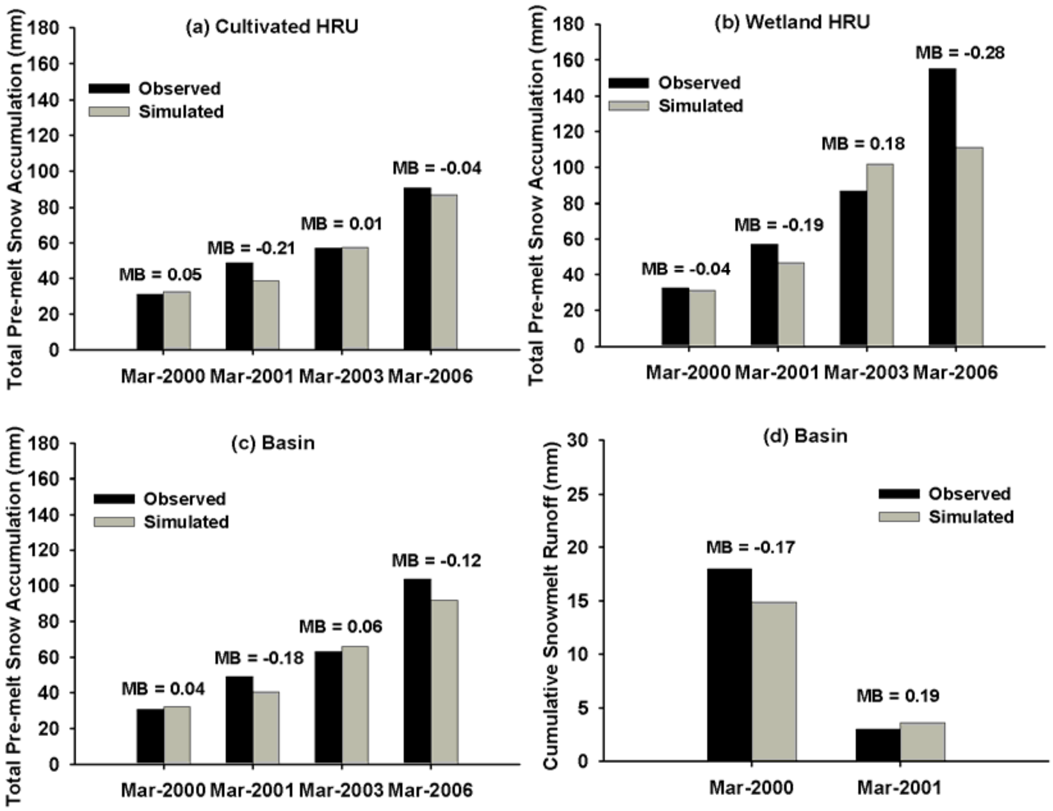

Figure 8. CRHM test of total pre-melt snow accumulation: (a) cultivated HRU, (b) wetland HRU, and (c) basin; (d) CRHM test of cumulative basin snowmelt runoff

observed total runoff from snowmelt at the end of March in 2000 and 2001 (Figure 8(d)). For the simulation of snowmelt runoff in end of March 2000, 14.9 mm surface runoff was estimated from the snowmelt at the Wetland 109 including the cultivated field (contributing area) and wetland (runoff accumulating area). This is comparable with the observed snowmelt runoff. Moderate value of Model Bias (MB) -0.17, represents an underestimation of 3 mm for the surface runoff from melting water. CRHM simulations estimated about 3.6 mm from the contributing area to the wetland at the end of March 2001; this is very comparable to the observed melt runoff. Both simulation and observation show that very low snowmelt runoff occurred in March 2001. More discussion on evaluation and application of CRHM is given by Fang and Pomeroy (2007) and Pomeroy *et al*. (2007b).

Drought Impacts Simulations

CRHM is able to analyze drought impacts on hydrological processes (Fang and Pomeroy, 2007). These impacts are both meteorological and through water storage and land use response to drought. Through the CRHM simulations, combined effects of changing meteorology and changing land cover and soil conditions during drought can be detected and examined. Figure 9 shows the impact of recent Canadian prairie drought on wetland

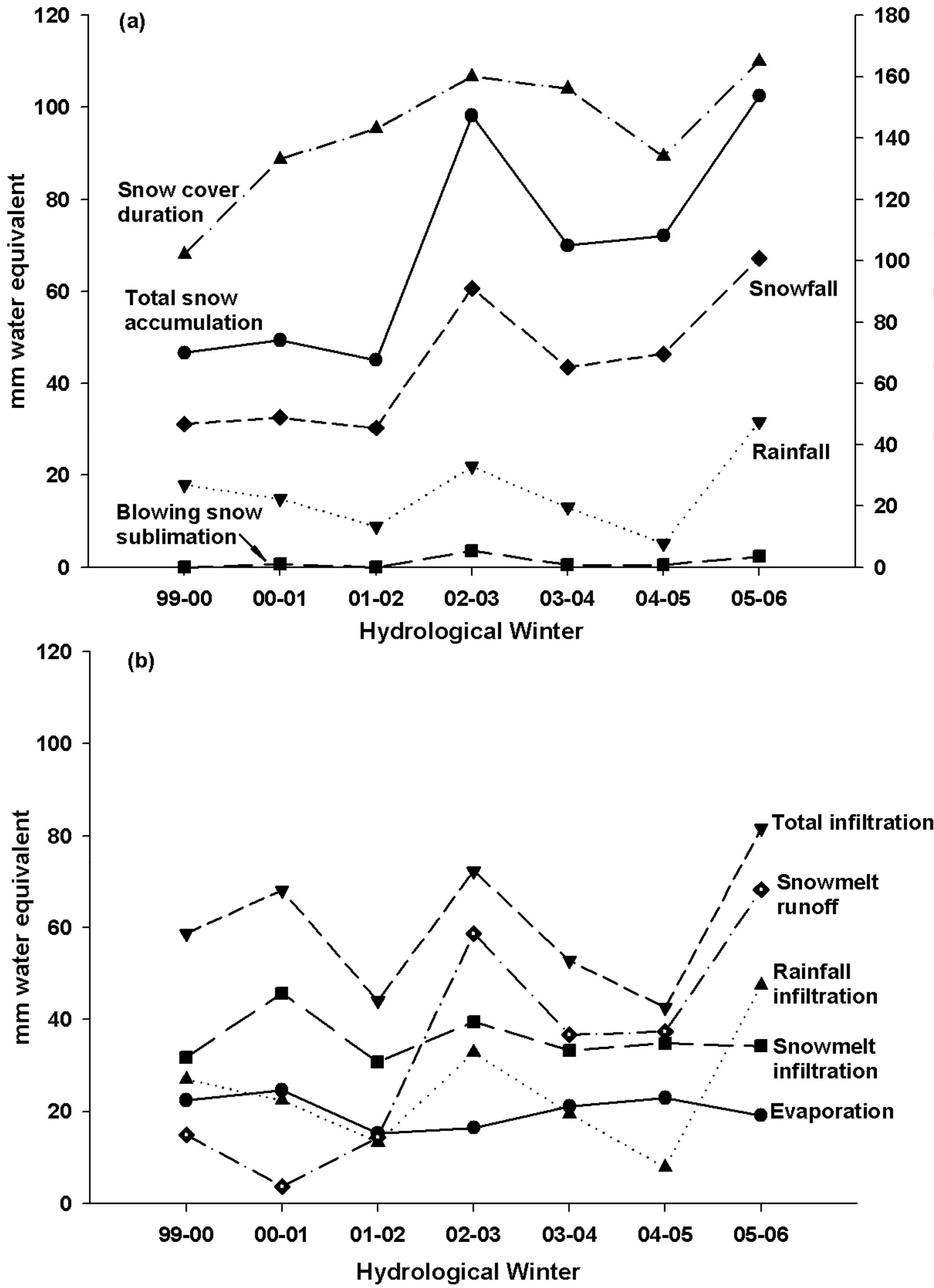

Figure 9. Hydrological processes for the basin of Wetland 109 during 1999-2006: (a) snowfall, rainfall, total snow accumulation, snow cover duration, blowing snow sublimation and (b) evaporation, rainfall infiltration, snowmelt infiltration, total infiltration, snowmelt runoff

winter hydrological processes. The CRHM simulation showed that cumulative snow accumulation was lower with shorter snow-covered seasons and slightly enhanced evaporation during the severe drought periods of 1999-2002; other processes (e.g. infiltration) did not show a particular trend because infiltration comprises both rainfall infiltration into unfrozen soils and snowmelt infiltration into frozen soils; both are very complex processes controlled by combinations of hydro-meteorological condition and soil status. Basin surface runoff from snowmelt was much lower during the severe drought periods, approximately 45 to 65 mm less compared to that in the normal periods of 2005-06. The low surface runoff is compared to the field observations as shown in Figure 3. Any confidence in the CRHM simulations for drought impact on the snowmelt runoff-related processes is due to he strong physical basis of modules that are key to the water balance; these modules take into consideration all of the snowmelt runoff processes that are relevant to the prairie wetland environment.

7. Hydrological Modelling Approach

A hydrological study to investigate the effects of depression (wetland) restoration on the surface runoff to Devils Lake was conducted by WEST Consultants, Inc. (WEST, 2001). The study was carried out in the upper basin of Devils Lake watershed, North Dakota. Various techniques – digital elevation model (DEM), National Wetlands Inventory (NWI) data, aerial photos were used to characterize, delineate the basin and to determine the location, area, volume of depressions, and a systematic classifying scheme was developed to categorize the depressions within the basin. A customized hydrological model – Pothole-River Network Model (PRINET) was developed from the HEC Hydrologic Modeling System (HEC-HMS) (Hydrologic Engineering Center, U.S. Army Corps of Engineers, 2001). The PRINET model determines the daily precipitation and evaporation for each sub-basin, calculates soil moisture, computes the runoff to the on- and off-river 'intact' depressions, and routes the excess runoff to on-river 'intact' depressions in the downstream sub-basins. Despite several drawbacks from this study – a lack of physical algorithms for soil moisture accounting and infiltration simulation, non-physically-based temperature-index snowmelt routine, and lack of field validation of depression classification, it might provide a useful framework for depression (wetland) classification and assessment of wetland restoration effects on the downstream runoff. This framework might form the basis of a modelling approach using CRHM.

Considering a number of techniques involved, there will be four procedures that are incorporated into the hydrological modelling approach for the prairie hydrology study as shown in Figure 10. The first procedure deals with geo-spatial information processing that defines basin and sub-basins and generates a classification of wetlands and catchments. This procedure serves as a hydrological model pre-processing purpose, providing spatial and physical features of the basin for subsequent hydrological modelling. The second procedure involves applying forcing meteorological data and field site hydrological parameters to the model CRHM. Various model outputs will include snow accumulation, snowmelt, infiltration, actual evaporation, soil moisture status, and streamflow, and field observations of snow surveys, snowmelt, and streamflow will be

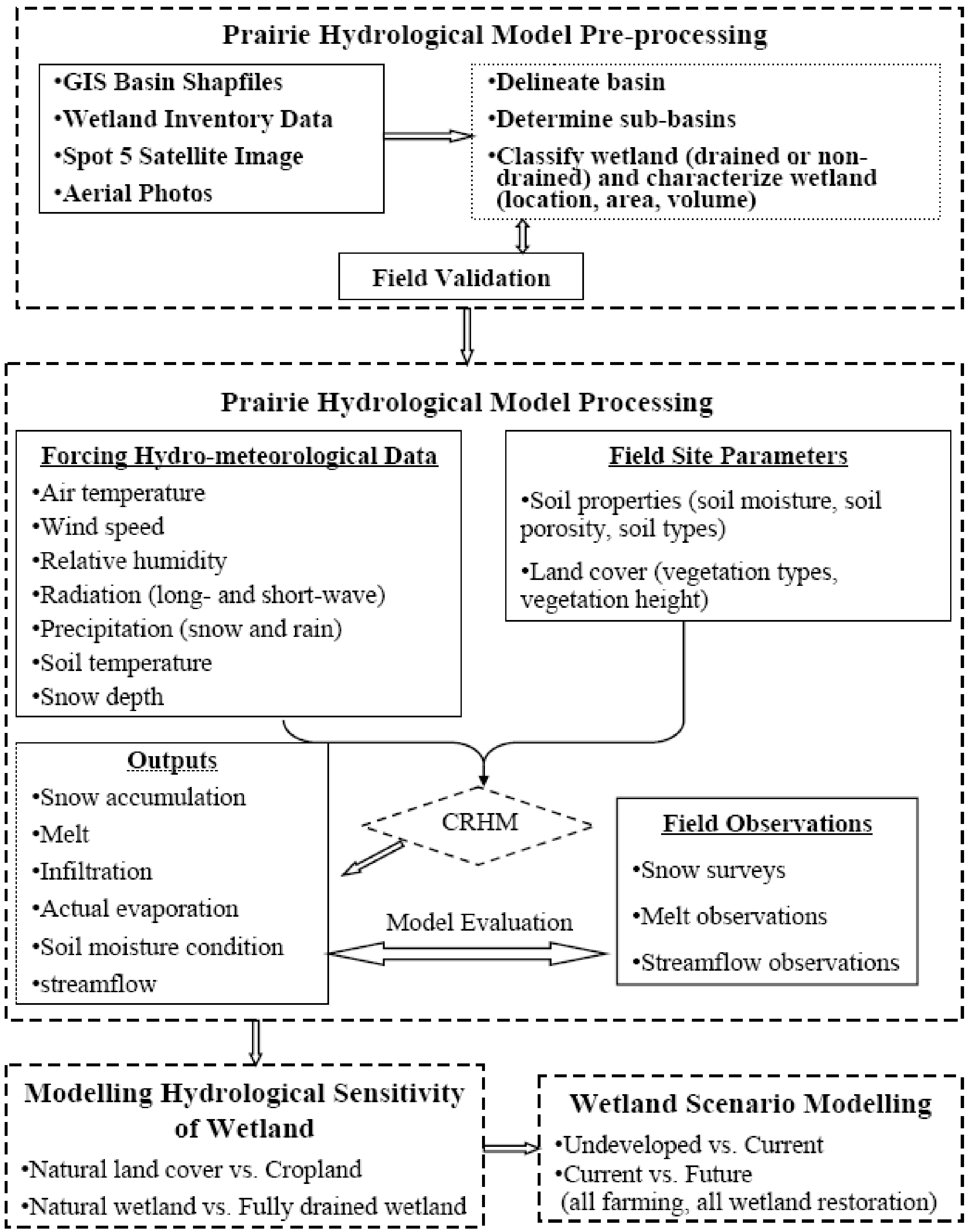

Figure 10. Modelling approach for prairie hydrology study.

compared to modelled results to evaluate model performance. The last two procedures are applications of the prairie hydrological model, starting with hydrological sensitivity analysis of the effects of land cover (wetland, natural upland or cropland) and drainage (natural or drained) on prairie hydrology (wetland water level, downstream flow). This is followed by scenario modelling, which assesses the response of wetland water level and downstream flow to various scenarios – undeveloped wetland condition, current wetland condition, and future condition. Throughout this study, it is expected to advance the understanding of hydrology in the prairie region, provide suggestions to farming, flood control, wetland management, and to facilitate the development of a proper prairie hydrological model for water resources assessments.

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APPENDIX A – Modelling Scheme for Prairie Hydrological Model