HYDROLOGICAL PROCESSES
Hydrol. Process. (2009)
Published online in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/hyp.7348

Modelling blowing snow redistribution to prairie wetlands

X. Fang* and J. W. Pomeroy
Centre for Hydrology, University of Saskatchewan, Saskatoon, Canada

Abstract:

Blowing snow transports and sublimates a substantial portion of the seasonal snowfall in the prairies of western Canada. Snow redistribution is an important feature of prairie hydrology as deep snowdrifts provide a source of meltwater to replenish ponds and generate streamflow in this dry region. The spatial distribution of snow water equivalent in the spring is therefore of great interest. A test of the distributed and aggregated modelling strategies for blowing snow transport and sublimation was conducted at the St. Denis National Wildlife Area in the rolling, internally drained prairie pothole region east of Saskatoon, Saskatchewan, Canada. A LiDAR-based DEM and aerial photograph-based vegetation cover map were available for this region. A coupled complex windflow and blowing snow model was run with 262,144 6 m × 6 m grid cells to produce spatially distributed estimates of seasonal blowing snow transport and sublimation. The calculation was then aggregated to seven landscape units that represented the major influences of surface roughness, topography and fetch on blowing snow transport and sublimation. Both the distributed and aggregated simulations predicted similar end-of-winter snow water equivalent with substantial redistribution of blowing snow from exposed sparsely vegetated sites across topographic drainage divides to the densely vegetated pothole wetlands. Both simulations also agreed well with snow survey observations. While the distributed calculations provide a fascinating and detailed visual image of the interaction of complex landscapes and blowing snow redistribution and sublimation, it is clear that blowing snow transport and sublimation calculations can be successfully aggregated to the spatial scale of the major landscape units in this environment. This means that meso and macroscale hydrological models can represent blowing snow redistribution successfully in the prairies.

KEY WORDS blowing snow; snow transport; snow sublimation; parameterization; hydrological modelling; prairie wetlands

Received 30 June 2008; Accepted 26 March 2009

INTRODUCTION

Snow is an important water resource on the Canadian Prairies because it preferentially generates runoff compared to rain. Gray and Landine (1988) noted that in west-central Saskatchewan approximately one-third of annual precipitation occurs as snowfall, which produces 80% or more of annual local surface runoff. Techniques to estimate snow accumulation and its variability need to take into account the governing processes and scale at which they operate. Pomeroy and Gray (1995) formulated three spatial scales describing the variability of snow accumulation: micro (10–100 m), meso (100 m–10 km), and macro (10–1000 km). Snow accumulation is highly heterogeneous at micro- and mesoscales in open environments, due to redistribution by blowing snow. Wind redistribution of snow is primarily from open, well exposed sites to sheltered or vegetated sites. Whilst other processes operate at mesoscales to cause snow accumulation variability such as mesoscale flows, elevational gradients in precipitation and lake effect snowfalls, these other processes are suppressed on the prairies due to generally level topography and frozen water bodies in winter.

There are two primary modes of movement involved in the transport of blowing snow–saltation and suspension (Pomeroy and Gray, 1995). Blowing snow transport forms snowdrifts, usually in depressions, 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 micro- and mesoscale heterogeneity in snow accumulation are 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 the seasonal snowfall on the Canadian Prairies (Pomeroy and Gray, 1995). Transport and sublimation of blowing snow result in significant ablation of the snowpack before spring melt occurs. Blowing snow can transport and sublimate as much as 75% of annual snowfall from open, exposed fallow fields in southern Saskatchewan (Pomeroy and Gray, 1995).

Topography and land cover roughness are the two major factors influencing blowing snow transport, because both induce variations in wind speed near the snow surface. In absence of exposed vegetation, a leeward slope has a much greater snow accumulation than does a windward slope (Steppuhn, 1981; Pomeroy and Gray, 1995). Lapen and Martz (1996) showed similar findings from extensive snow surveys in the region, suggesting that the spatial distribution of snow depth in a prairie agricultural landscape is strongly influenced by the

* Correspondence to: X. Fang, Centre for Hydrology, University of Saskatchewan, 117 Science Place, Saskatoon, Sask. S7N 5C8. Canada. E-mail: xif382@mail.usask.ca

orientation of slopes with respect to the primary directions of wind transport, and their relative position to other topographic features. Different land covers impose variations in surface roughness, which in turn causes wind speed and snow accumulation to change. Pomeroy *et al.* (1990) found that wheat stubble fields in southern Saskatchewan had substantially smaller losses to blowing snow compared to fallow fields because the exposed stubble reduced the shear stress imposed by the wind on the snow surface. At Bad Lake, Saskatchewan, blowing snow accumulation in the depressions with tall shrubs enhanced snowpacks by approximately 50–100% of seasonal snowfall (Pomeroy *et al.*, 1998).

The Prairie Blowing Snow Model (PBSM) developed by Pomeroy (1989) assembles physically based wind transport and sublimation algorithms to solve a mass continuity equation at the surface and hence estimate seasonal snow accumulation on the Canadian Prairies (Pomeroy *et al.*, 1993). Early versions of PBSM had simple representations of the effects of varying fetch and land cover. Upscaling of blowing snow transport and sublimation estimations to mesoscales was related to the variability of blowing snow transport and sublimation over open snow areas, the increase in transport and sublimation with fetch and the influence of exposed vegetation on the shear stress available to drive transport (Pomeroy and Li, 2000). Pomeroy *et al.* (1997) developed a simple scheme to address the calculation of areal snow mass balance in a basin based on monthly climatological expressions of blowing snow transport and sublimation. These expressions were not directly applicable in other atmospheric and hydrological models due to their empirical nature and monthly time step, but the distribution methodology provided the basis to distribute blowing snow over landscape and was further evaluated by Essery and Pomeroy (2004).

Spatially distributed blowing snow models have been developed to simulate blowing snow fluxes over complex landscapes. These models represent topography and vegetation cover by high-resolution grids and redistribute snow along windflow vectors whilst calculating sublimation losses (Liston and Sturm, 1998; Essery *et al.*, 1999). The windflow model of Mason and Sykes (1979) is physically based. This model is used to estimate the wind speed and direction variation due to local topography in the blowing snow simulation of Essery *et al.* (1999), which distributes a parametric version of PBSM, called the Distributed Blowing Snow Model (DBSM). DBSM applied at a point produces approximately the same output as PBSM except for a slightly simpler treatment of vegetation effects on surface wind speed.

The Canadian Prairies contain many millions of poorly drained topographic depressions as a remnant of relatively recent glacial erosion and deposition. These depressions are referred to as 'potholes' in the USA and 'sloughs' in western Canada (Woo and Rowsell, 1993) and they form wetlands. Wetlands are important elements in prairie surface hydrology as they have great runoff retention capacity (Hayashi *et al.*, 2003), and they are also critical habitats for waterfowl and other wildlife. These wetlands have different storage capacities and many of them are internally drained, and so, do not normally connect with one another except during the spring freshet after exceptionally wet and high snowfall conditions; hence their connectivity and storage capacity are critical factors in determining surface runoff (Godwin and Martin, 1975; Spence, 2007). Snow drifts form in prairie wetland areas as a result of redistribution during winter blowing snow events. In the subsequent spring melt periods, high water levels in wetlands, and sometimes local runoff, develop from the rapid release of water from snowpacks at a time of limited or restricted infiltration and low evaporation rates (Hayashi *et al.*, 2003; van der Kamp *et al.*, 2003). Wetland water levels drop during the summer as rainfall and residual spring snowmelt water are normally consumed by evaporation (Woo and Rowsell, 1993; Conly and van der Kamp, 2001).

Due to the importance of snow accumulation in the water balance of prairie wetlands, it is vital to estimate snow accumulation correctly. However, there is a concern that fine-scale distributed simulations of blowing snow will not be practical for wetland hydrology modelling over such a large region due to computational costs (Essery and Pomeroy, 2004) and difficulties in parameterization at fine scales. Thus, it is considered necessary to have an aggregated approach with an appropriate level of accuracy for estimating end-of-winter snow redistribution to wetlands for large regions. The objectives of this study are therefore to:

1. Determine whether winter blowing snow to prairie wetlands can be estimated accurately using blowing snow models when compared to field observations.
2. Examine whether accurate information on spring snow accumulation in prairie wetlands can be determined from either spatially aggregated or spatially distributed modelling approaches.
3. Recommend an appropriate model complexity for physically based models of blowing snow redistribution amongst landscape-based land units.

STUDY AREA

The study was conducted at the St. Denis National Wildlife Area (SDNWA). The SDNWA (52°02′N, 106°06′W, 545–560 m a.s.l.) is a small basin (3·85 km^2), located in south-central Saskatchewan, Canada (Figure 1), in a moderately rolling landscape with slope ranging from 10 to 15% (van der Kamp *et al.*, 2003). The area is dominated by small wetlands with poorly-developed drainage, clay soils and glacial till substrate (Hayashi *et al.*, 1998). The SDNWA has three major land uses: native grassland, brome grassland, and cultivated land. Saskatoon Airport, located about 40 km west of the SDNWA, has a 2 °C annual air temperature, with −19 °C as the January mean temperature and 18 °C as the July mean temperature; the 30-year (1967–1996) mean

DOI: 10.1002/hyp

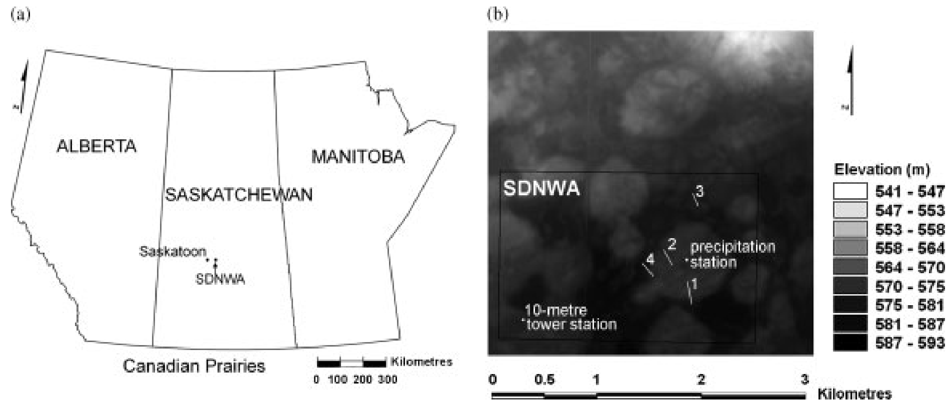

Figure 1. Study area: (a) SDNWA, Saskatchewan and (b) plan of field observations at SDNWA on a digital elevation model (DEM) (dark solid line indicates SDNWA boundary and while solid lines indicates field observation transects; white points denote meteorological stations)

annual precipitation in Saskatoon is 358 mm with 74 mm of snowfall occurring from November to April (van der Kamp *et al*., 2003). Snowfall generally starts in November, and several snowmelt runoff events occur in early spring between March and April in the area (van der Kamp *et al*., 2003).

STUDY METHODS

Snow surveys

Snow surveys were conducted during January–April 2006, along four transects shown in Figure 1(b). The surveys were to capture evolution of snow water equivalent (SWE) throughout the winter. In total, there were 138 sampling points at 5-m intervals. At each point, snow depth was recorded by inserting a 1-m ruler into the snowpack. At every fifth depth measurement point, a 7-cm-diameter ESC-30 snow tube was inserted into the snowpack to extract a sample of snow which was then weighed to determine snow density. Average snow densities for each transect were calculated and used with the point depth measurements to determine observed SWE for each depth point (Pomeroy and Gray, 1995). A brief description of the snow survey transects is presented in Table I.

The SDNWA was broken down into landscape units termed ‘hydrological response units’ (HRUs) following the guideline for stratification of landscape for prairie snow accumulation by Steppuhn and Dyck (1974). HRUs are the largest landscape units having definable hydrological characteristics, in that they can be described by unique sets of parameters, variables and fluxes (Pomeroy *et al*., 2007b). HRUs can be represented by the average SWE from portions of transects and the average topography, location and vegetation characteristics.

Table I. Description of snow accumulation survey field transects at SDNWA

Transect #	Length (m)	Sampling points	Depth sampling interval (m)	Description
1	220	45	5	Rolling stubble field
2	170	35	5	Relatively flat stubble field crossing a small wetland
3	120	25	5	Relatively flat grassland
4	160	33	5	Rolling field with transition between grassland and stubble field

Meteorological data

Meteorological measurements were carried out from 2005 to 2006 at the precipitation station and the 10-m tower station shown in Figure 1(b). The stations included a hygrothermometer, anemometer, Alter-shielded Geonor precipitation gauge, tipping bucket rain gauge and blowing snow particle detector (Brown and Pomeroy, 1989). Hourly measurements of air temperature, relative humidity, 10-m wind speed and wind direction, snowfall and rainfall were collected in the winter of 2005–2006 (Figure 2). Snowfall was corrected for wind-undercatch using the algorithm of MacDonald and Pomeroy (2008). The meteorological observations were used to drive blowing snow models for the period from 31 October 2005 to 27 March 2006.

Digital elevation model and aerial photograph-based vegetation height map

Grid cells of a LiDAR-derived digital elevation model (DEM) and vegetation height that are used by the Distributed Blowing Snow Model (DBSM) (Essery *et al*., 1999) were collected. The LiDAR campaign was carried out by Applied Geomatics Research Group (AGRG) of Nova Scotia Community College and C-CLEAR programme on 9 August 2005. Optech’s proprietary REALM software was used to process all data and ‘flat terrain’, and ‘dense vegetation’ options were applied to separate ground and non-ground features (Töyrä, 2005). Further filtering was conducted by Environment Canada, and LiDAR explorer for ArcGIS was used to separate bare ground layer and vegetation layer based on a 3×3 kernel size and 0·2 m Z tolerance. This resulted in nine LiDAR tiles covering SDNWA and the surrounding area. All data in the nine LiDAR tiles were interpolated into a DEM using Inverse Distance Weighted (IDW) algorithm

DOI: 10.1002/hyp

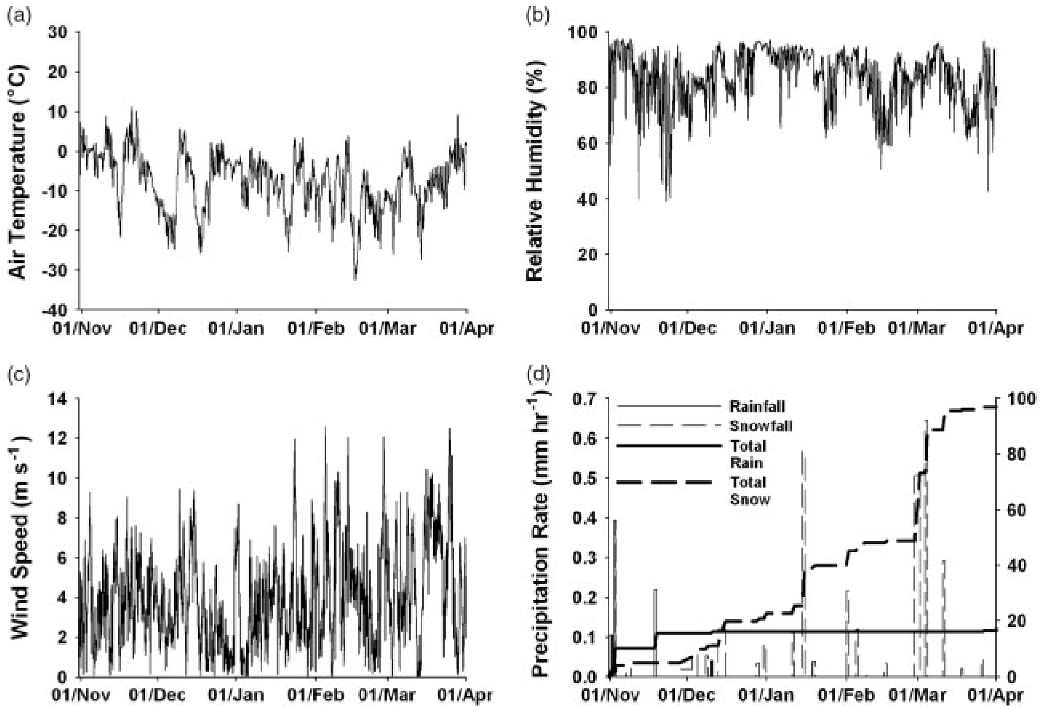

Figure 2. Hourly observations of (a) air temperature (b) relative humidity (c) wind speed and (d) precipitation during period from 31 October 2005 to 30 April 2006

in ArcGIS (Töyrä, 2005). This results in the total number of 6201 × 6201 grid cells with a cell size of 0·5 × 0·5 m for the DEM. The fast Fourier transform routine in the windflow model in DBSM requires a power-of-two number of cells in each direction of the DEM. Thus, a grid cell resample routine was conducted in ArcView GIS, resulting in 512 × 512 grid cells with a cell size of 6 × 6 m.

For the vegetation height grid, aerial photos and site plan along with vegetation height were used. A detailed vegetation survey was conducted from 24 to 26 October 2005. Vegetation height was measured by ruler. Eight groups of vegetation were classified based on the height and type: fallow (bare ground), short stubble (cereal grains), tall stubble (cereal grains), short grass, tall grass, shrubs, short trees, and tall trees along with non-vegetated features (roads and frozen water bodies treated as zero height). Vegetation-height polygons were created and delineated in ArcView GIS, resulting in an ArcView GIS shape file containing vegetation height. This shape file was converted into a grid file comprising 512 × 512 grid cells with a cell size of 6 × 6 m. Both grids of DEM and vegetation height are shown in Figure 3.

Modelling blowing snow

Spatially distributed and spatially aggregated approaches to modelling blowing snow were used to estimate SWE over the period from 31 October 2005 to 27 March 2006. Both approaches used models with similar physics (PBSM, DBSM) that differed in the spatial scale at which calculations were carried out. Calculations in the spatially distributed approach were conducted on small-sized grid cells, whereas calculations in the spatially aggregated approach were based on large-sized HRUs.

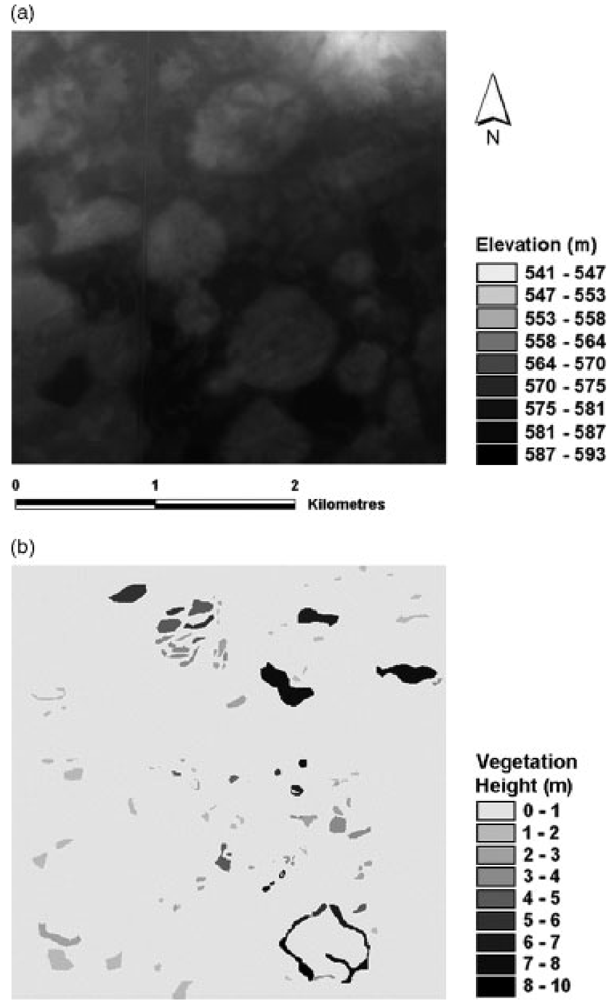

Figure 3. Gridded maps of (a) LiDAR-based digital elevation model (DEM) and (b) aerial photograph-based vegetation height (note individual grid elements cannot be seen at this scale)

DOI: 10.1002/hyp

Spatially distributed approach. A portable version of the DBSM (Essery *et al.*, 1999) was used in this approach. It is a distributed parametric version of the PBSM (Pomeroy, 1989; Pomeroy *et al.*, 1993), which uses wind fields simulated by the windflow model of Mason and Sykes (1979), (MS). The MS windflow component is a computational routine for estimating windflow over three-dimensional topography, which is extended from the two-dimensional theory of Jackson and Hunt (1975) for turbulent flow over a shallow hill. The MS windflow component is based on Fourier transform techniques and assumes neutral thermal stratification and uniform surface roughness within the simulation region. MS calculates normalized westerly and southerly wind components along with normalized wind speed due to changing topography. The normalized wind speed adjusts the local wind speed to estimate changes in wind speed due to variable topography. This then has important controls on blowing snow fluxes in the rest of DBSM. The calculation in the portable DBSM was conducted on 262,144 grid cells of DEM and vegetation height with 6-m spatial resolution. On each grid cell, the flux of snow mass was adjusted for transport and sublimation fluxes by the continuity equation:

$$\frac{dSWE}{dt} = S_f - q_s - \nabla \cdot q_T \quad (1)$$

where SWE (mm or kg m^{-2}) is simulated SWE, S_f (kg m^{-2} s^{-1}) is the snowfall rate, q_s (kg m^{-2} s^{-1}) is the sublimation rate, and $\nabla \cdot q_T$ (kg m^{-2} s^{-1}) is the horizontal divergence of transport (Figure 4).

Spatially aggregated approach. The PBSM of Pomeroy and Li (2000) was coupled in the Cold Regions Hydrological Model Platform (CRHM) with a simple windflow model to estimate the redistribution of snow by wind due to changing local topography and surface roughness. The windflow model is a parametric version of the MS windflow model (Walmsley *et al.*, 1989). The parametric windflow model considers the effect of small-scale topographical variations on wind speed and adjusts the wind speed accordingly, such that wind speeds over hilltops are greater than those over depressions or flat terrain. The adjusted wind speeds from the windflow model are used in PBSM, which uses physically based algorithms to estimate snow accumulation based on Eq. 1. The calculation is applied to control volumes corresponding to landscape HRUs (Figure 5(a)). To create these HRUs, landscapes are divided into source and sink regions based upon aerodynamic considerations as governed by topography and vegetation (Figure 5(b)). Source areas include upland fields and hilltops, and sink areas include aspen bluffs, ravines, streams and wetlands. HRUs for blowing snow do not necessarily have a precise geographical location but they do have an aerodynamic sequence and proximity to each other which can be characterized for the prairies. The characteristic sequence is from uplands of cultivated

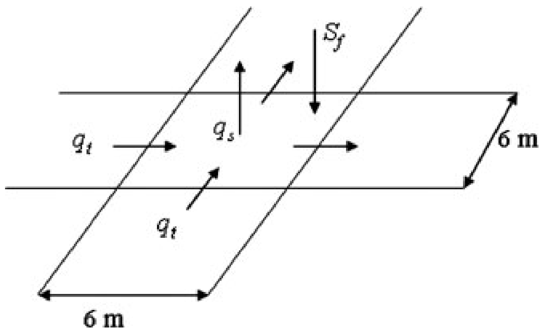

Figure 4. Schematics of the snow mass calculation on a grid in the spatially distributed approach. Arrows indicate fluxes of snow mass

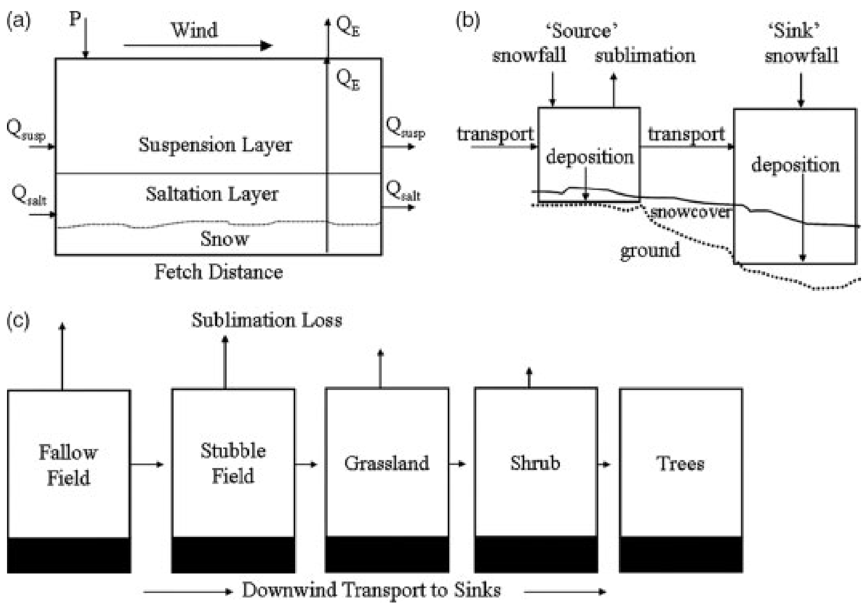

Figure 5. Schematics of the snow mass calculation in the spatially aggregated approach, (a) cross-sectional view of the control volume for blowing snow mass fluxes, (b) source and sink landscape units control volumes and (c) characteristic sequence of blowing snow HRU control volumes for the prairies

fields to steep hillsides of grass and shrub to lowlands of grass, shrub, trees and wetlands in depressions as shown in Figure 5(c). In the spatially aggregated approach, 19 initial HRUs based upon combinations of topographic exposure and vegetation were eventually simplified to seven HRUs based on distinct snow accumulation and transport characteristics after initial model tests. The final number of HRUs was also guided by the classification of prairie snow accumulation from Steppuhn and Dyck (1974). These seven HRUs are shown in Figure 6 and are considered to be the major landscape units at SDNWA with respect to redistribution of snow by wind. Table II also shows key parameters for the seven HRUs in the PBSM calculations. The area of each HRU was determined by ground survey; vegetation height was assigned based on field surveys of vegetation and aerial photography. Blowing snow fetch distance is the upwind distance without disruption to the flow of snow and was assumed to be equal to the PBSM minimum of 300 m, due to the rolling terrain characteristics of the field site.

To parameterize the flow of snow from HRU to HRU, PBSM as implemented in CRHM (Pomeroy *et al.*, 2007b) has a unique parameterization scheme employing the concept of cascading HRUs. Snow is redistributed amongst HRUs based upon snow transport calculations, HRU size, vegetation characteristics, and a distribution factor, D_p. More specifically, PBSM uses the conceptual distribution function to allocate blowing snow transport from aerodynamically smoother (or windier) HRUs to aerodynamically rougher (or calmer) ones within a

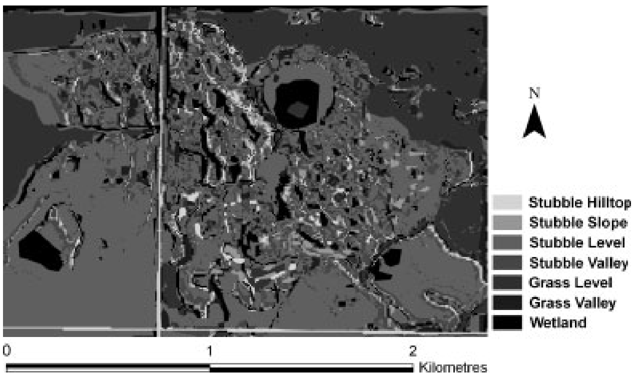

Figure 6. Map of seven HRUs at SDNWA used in the spatially aggregated approach

Table II. Characteristics of parameters for HRUs in the spatially aggregated approach

HRUs name	Area (km^2)	Vegetation height (m)	Blowing snow fetch distance (m)	Blowing snow distribution parameter
Stubble hilltop	0·5	0·05	300	1
Stubble slope	0·4	0·12	300	1
Stubble level	1	0·15	300	1
Stubble valley	0·5	0·2	300	3
Grass level	1	0·5	300	2
Grass valley	0·3	0·5	300	5
Wetland	0·15	5	300	10

basin. At the end of a day, when blowing snow transport has occurred, snow transported from HRUs with low aerodynamic roughness (or higher wind speed) is distributed over the HRUs with greater roughness (or lower wind speed) in the basin according to the fractions specified by D_P. Snow transport can be permitted to enter the 'basin' as the flux from the smoothest HRU and to leave the basin when the flux from the roughest HRU is greater than zero. When HRUs with lower vegetative height (or higher wind speed) fill until vegetation is buried, the excess is distributed over the remaining unfilled HRUs. For example, if the transport q_T out of some HRU is redistributed over the rougher HRUs 'A', 'B' and 'C' with their D_p having values of D_p(a), D_p(b) and D_p(c), respectively, the snow transport would be distributed as: $D_p(a)\times q_T/(D_p(a)+D_p(b)+D_p(c))$ to HRU A; $D_p(b)\times q_T/(D_p(a)+D_p(b)+D_p(c))$ to HRU B and $D_p(c)\times q_T/(D_p(a)+D_p(b)+D_p(c))$ to HRU C. With the concept of geographical proximity and connectivity of HRUs to each other in a basin based on HRU aerodynamic roughness, trial-and-error runs were employed to determine the values of D_p. These trial-and-error runs were guided by the relative size and spatial arrangement of HRUs. In Table II, a D_p value of 1 was assigned to stubble hilltop, stubble slope, and stubble level HRUs because these HRUs have low aerodynamic roughness (or high wind speeds), and are typically blowing snow 'sources'. In contrast, higher values of D_p were assigned to grass valley and wetland HRUs because they are blowing snow 'sinks' with high aerodynamic roughness (or low wind speeds). The assignment of D_p values is not a formalized process but remains a sufficiently simple and robust method for prairie environments. A more objective method of distributing flows amongst HRUs, suitable for more complex alpine environments, is presented in a companion paper in this issue (MacDonald *et al.*, 2009).

Blowing snow modelling approach comparison

The simulated end-of-winter SWE cells from the spatially distributed approach were compared to the observed end-of-winter SWE points from the field transects. In addition, observed pre-melt SWE from the four field transects was stratified by HRU to provide average HRU observations for comparison with an end-of-winter modelled SWE in the HRU. Without calibration of model parameters, average values of simulated pre-melt SWE on these seven HRUs from both approaches were compared with the averaged observed values on the corresponding HRUs during the period from 3 January to 27 March 2006.

To evaluate the performance of different modelling approaches, three statistical measures, Root Mean Square Difference (RMSD), Nash-Sutcliffe coefficient (NS) (Nash and Sutcliffe 1970) and Model Bias (MB) were calculated as:

$$RMSD = \frac{1}{n}\sqrt{\sum (SWE_s - SWE_o)^2} \qquad (2)$$

DOI: 10.1002/hyp

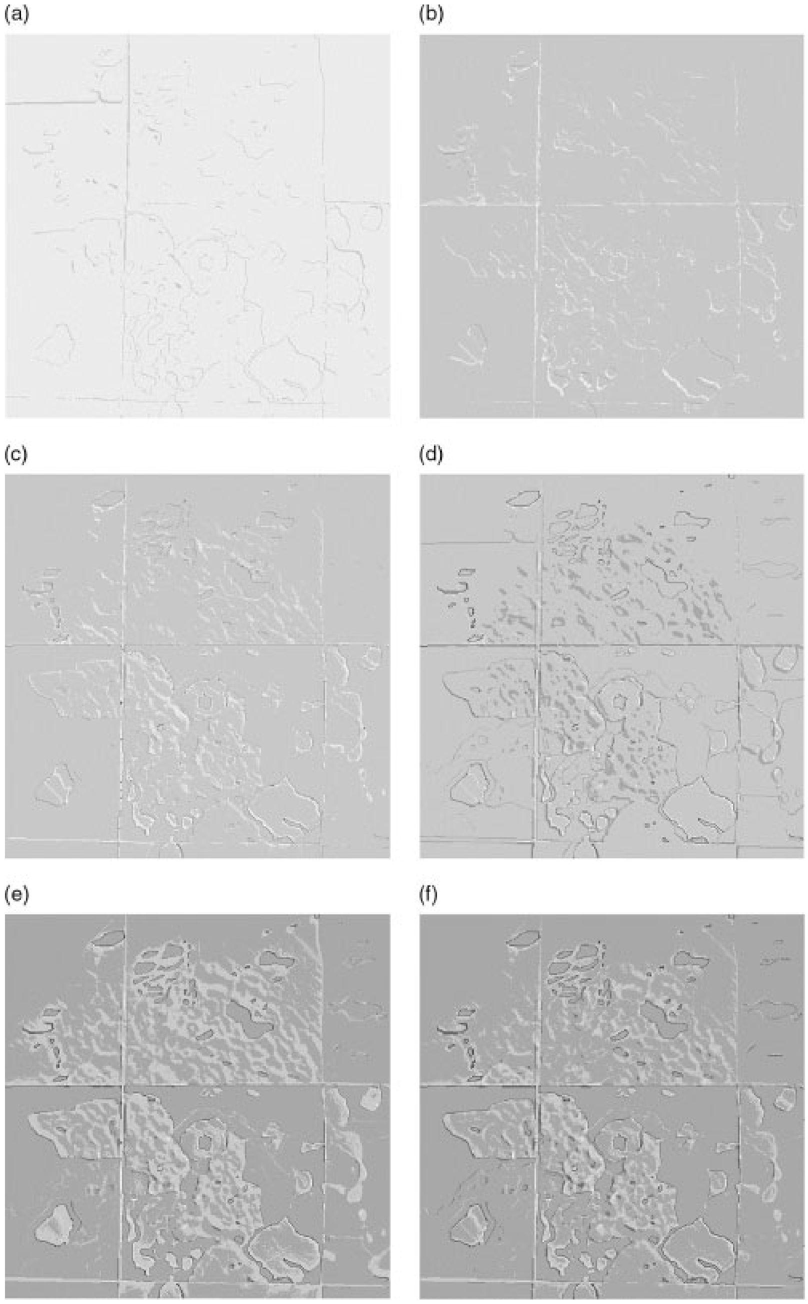

Figure 7a. Simulated evolution of pre-melt SWE from the spatially distributed approach for the winter of 2006 (a) 3 January, (b) 19 January, (c) 8 February, (d) 3 March, (e) 8 March, (f) 15 March, (g) 21 March, (h) 23 March and (i) 27 March

$$NS = 1 - \frac{\sum(SWE_o - SWE_s)^2}{\sum(SWE_o - \overline{SWE_o})^2} \quad (3)$$

$$MB = \frac{\sum SWE_s}{\sum SWE_o} - 1 \quad (4)$$

where n is number of samples, SWE_o, SWE_s, and $\overline{SWE_o}$ are the observed, simulated, and mean of the observed SWE, respectively. The RMSD is a weighted measure of the difference between observed and predicted SWE and has the same units as the observed and predicted SWE. A NS measures the model efficiency with a value equal to 1 implying that model perfectly predicts pre-melt SWE with respect to observations. A value equal to zero indicates that estimated values are not different from the average of observed values. 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). A positive value and a negative value of MB indicate overprediction and underprediction by the model, respectively.

RESULTS

Pre-melt snow accumulation

The simulations of pre-melt SWE from both spatially distributed and spatially aggregated approaches were run

DOI: 10.1002/hyp

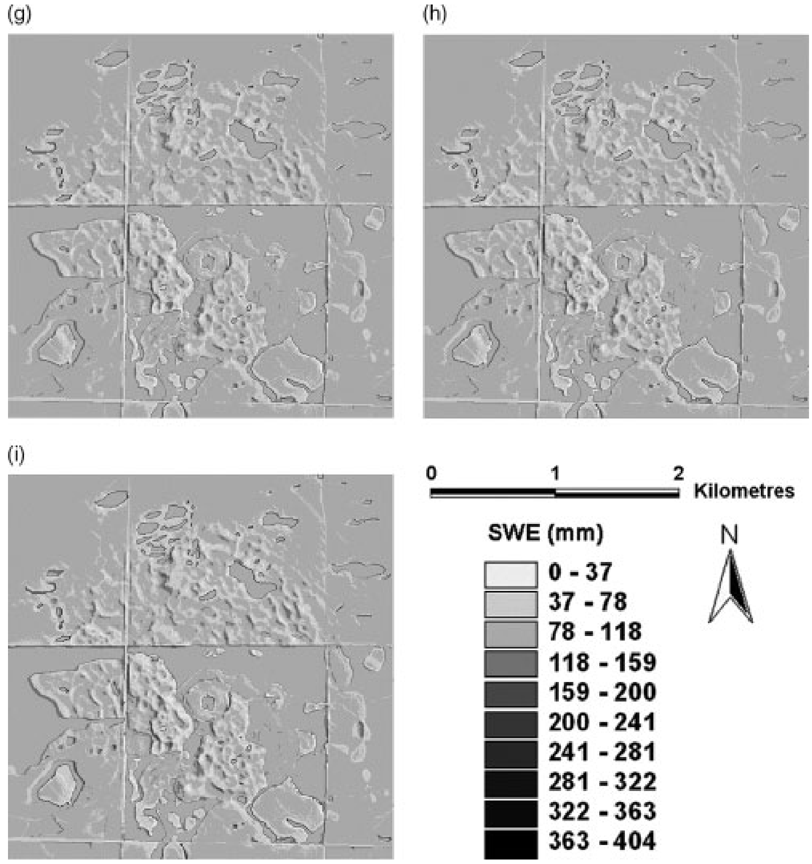

Figure 7b. (*Continued*)

for the 147-day period from 31 October 2005 to 27 March 2006. Cumulative snowfall was approximately 97 mm during this period. The areal distribution of snow accumulation for nine dates before melt was simulated from the spatially distributed approach (Figure 7). The Figure shows how snow accumulation changed from 3 January to 27 March 2006; the areal snow accumulation evolved rapidly after the three major snowfalls (24 mm on 28 February and 1 March, 15 mm on 4 March, and 8 mm on 11 March). Figure 7 also illustrates the spatial pattern of snow accumulation due to redistribution to depressions such as roadside ditches and to aerodynamically rough areas such as wetlands with tall vegetation. From the spatially aggregated approach, the spatial extent of the pre-melt SWE for the same nine dates was mapped on the seven HRUs at SDNWA and is shown in Figure 8. Figure 8 demonstrates a similar sequence of snow accumulation development to the distributed approach shown in Figure 7; snow was redistributed to and accumulated in areas with higher aerodynamic roughness such as grassland valleys and wetland HRUs. A time series of pre-melt SWE on the seven HRUs simulated from the spatially aggregated approach is shown in Figure 9; it demonstrates that taller vegetation (e.g. grassland and wetland) had greater snow accumulation than did shorter vegetation (e.g. stubble fields). More snow accumulated in valleys than on hilltops and slopes because hilltops and slopes have greater wind exposure, resulting in snow being drifted from these areas to valleys. At the end of the pre-melt period, about 148 mm and 237 mm of SWE were on 'grass valley' and 'wetland' areas, which was about four and six times that of the snow accumulation to the 'stubble hilltop' (i.e. 40 mm SWE), respectively. The 'stubble valley' area cumulated about 99 mm SWE, more than twice that of the 'stubble hilltop'.

Comparison between the distributed model and snow survey transects

The end-of-winter snow accumulation on 27 March 2006 simulated from the spatially distributed approach as simulated SWE on the four snow survey transects was compared to the observed SWE on these transects (Figure 10). Both model and observations show variability but no trend over distance for transects that sample a specific landscape type; that is, transect 1 (rolling stubble field) and transect 3 (level grassland). However, when transects cross a land cover type there are substantial differences in SWE in both simulations and observations; for instance in the middle section of transect 2 where land cover changes from level stubble field to wetland and in the middle section of transect 4 where land cover switches from stubble to grassland. The MB was calculated to quantify the difference between the simulated and observed SWE. The spatially distributed approach generally had reasonable estimations of the end-of-winter SWE; MB ranged from 0·19 to 0·31 for the simulated SWE on the four snow survey transects, indicating that

DOI: 10.1002/hyp

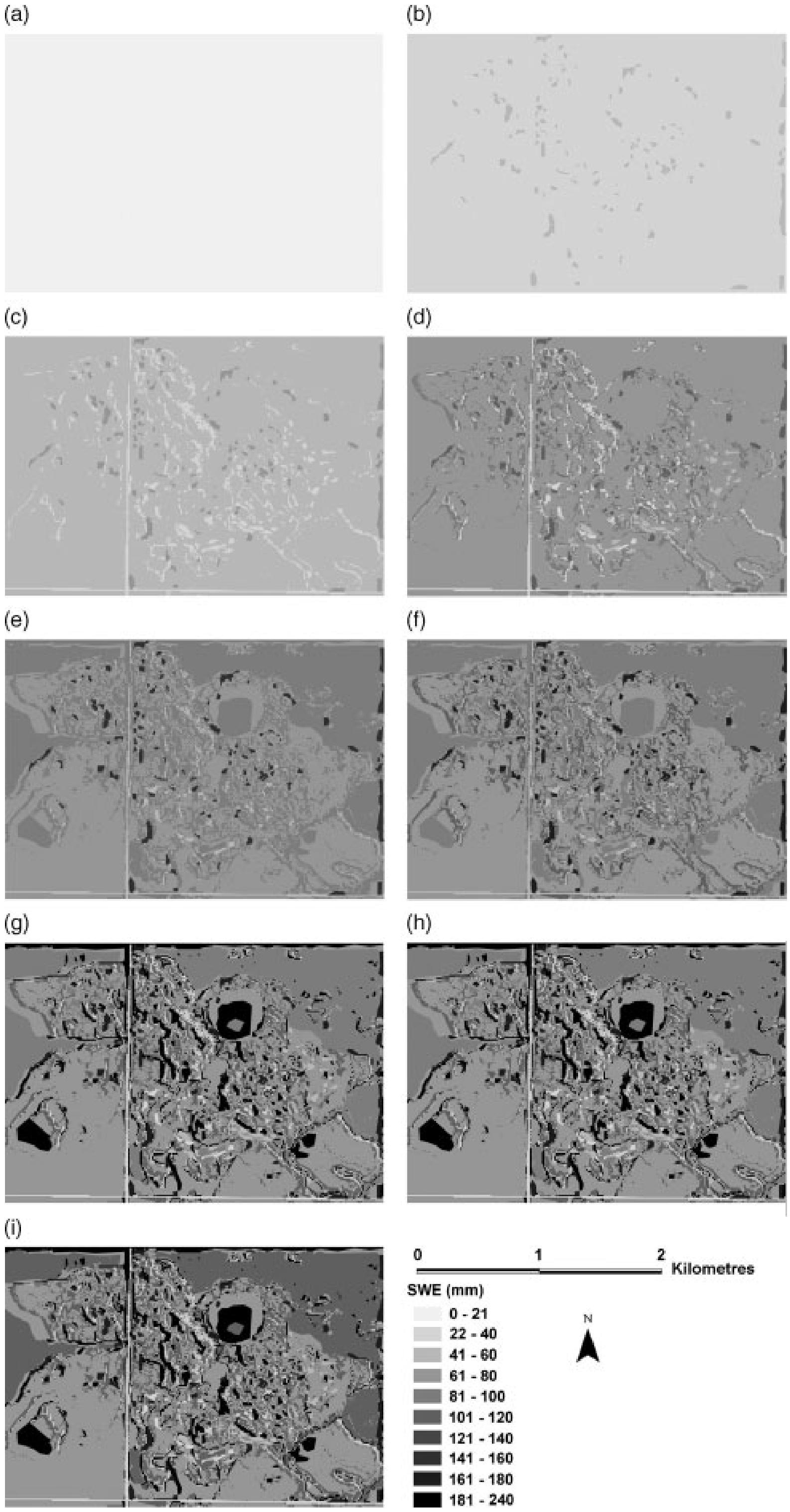

Figure 8. Spatial extent of simulated evolution of pre-melt SWE from the spatially aggregated approach for the winter of 2006 (a) 3 January, (b) 19 January, (c) 8 February, (d) 3 March, (e) 8 March, (f) 15 March, (g) 21 March, (h) 23 March and (i) 27 March

the spatially distributed simulation overestimated total transect SWE by 19–31%.

Comparison of distributed and aggregated models and snow surveys on the HRUs

Field observations of SWE were assigned to the seven HRUs and compared to simulated snow accumulation for the pre-melt period from January to March 2006 from both the spatially distributed and spatially aggregated modelling approaches (Figure 11 and Figure 12). These comparisons demonstrate that both approaches can simulate snow accumulations that are fairly close to observations at the landscape scale, but that the performance of each approach varied with landscape type.

To quantify the difference and performance of the two model approaches in predicting snow accumulation, the

DOI: 10.1002/hyp

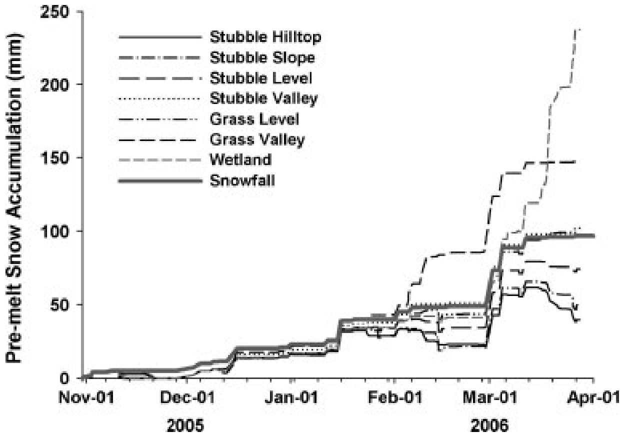

Figure 9. Temporal sequence of simulated pre-melt snow accumulation evolution from the spatially aggregated approach

RMSD, NS, and MB were computed for the seven HRUs (Table III). Both approaches had quite similar RMSD and NS for five HRUs ('stubble slope', 'stubble level', 'stubble valley', 'grass level' and 'grass valley'), with RMSD ranging from 2·06 mm to 5·91 mm and NS ranging from 0·68 to 0·91. This indicates that both approaches simulated the development and timing of snowcovers in winter period fairly well with relatively small differences with observations for these HRUs. For the 'stubble hilltop' HRU, a negative value of NS for the spatially distributed approach and a positive value of NS for the spatially aggregated approach implies that the spatially aggregated approach performs much better in predicting snowcover evolution here. For the 'wetland' HRU, values of NS are 0·84 and 0·22 for the spatially distributed and spatially aggregated approaches, respectively, indicating that the spatially distributed approach has better prediction power in the timing of SWE development.

The end-of-winter cumulative snow accumulation as of 27 March 2006 is shown in Figure 12. Both approaches

Table III. Root Mean Square Difference (RMSD) and Nash-Sutcliffe coefficient (NS) for the comparison of the simulated pre-melt snow accumulation from the spatially distributed (*Dis.*) and spatially aggregated (*Agg.*) approaches to observed pre-melt snow accumulation on the HRUs during period from January to March 2006; Model Bias (MB) for the comparison of the simulated total pre-melt snow accumulation from both modelling approaches to the observed total pre-melt snow accumulation on 27 March 2006

HRUs Name	RMSD (mm)		NS		MB	
	Dis.	*Agg.*	*Dis.*	*Agg.*	*Dis.*	*Agg.*
Stubble hilltop	5·57	3·74	−0·98	0·70	0·50	−0·37
Stubble slope	3·18	4·77	0·72	0·68	0·31	−0·36
Stubble level	2·95	2·06	0·81	0·85	0·26	−0·11
Stubble valley	3·29	2·25	0·87	0·86	−0·02	0·05
Grass level	3·19	4·36	0·86	0·86	0·16	−0·003
Grass valley	3·44	5·91	0·91	0·89	0·07	0·23
Wetland	9·58	30·61	0·84	0·22	−0·09	−0·06

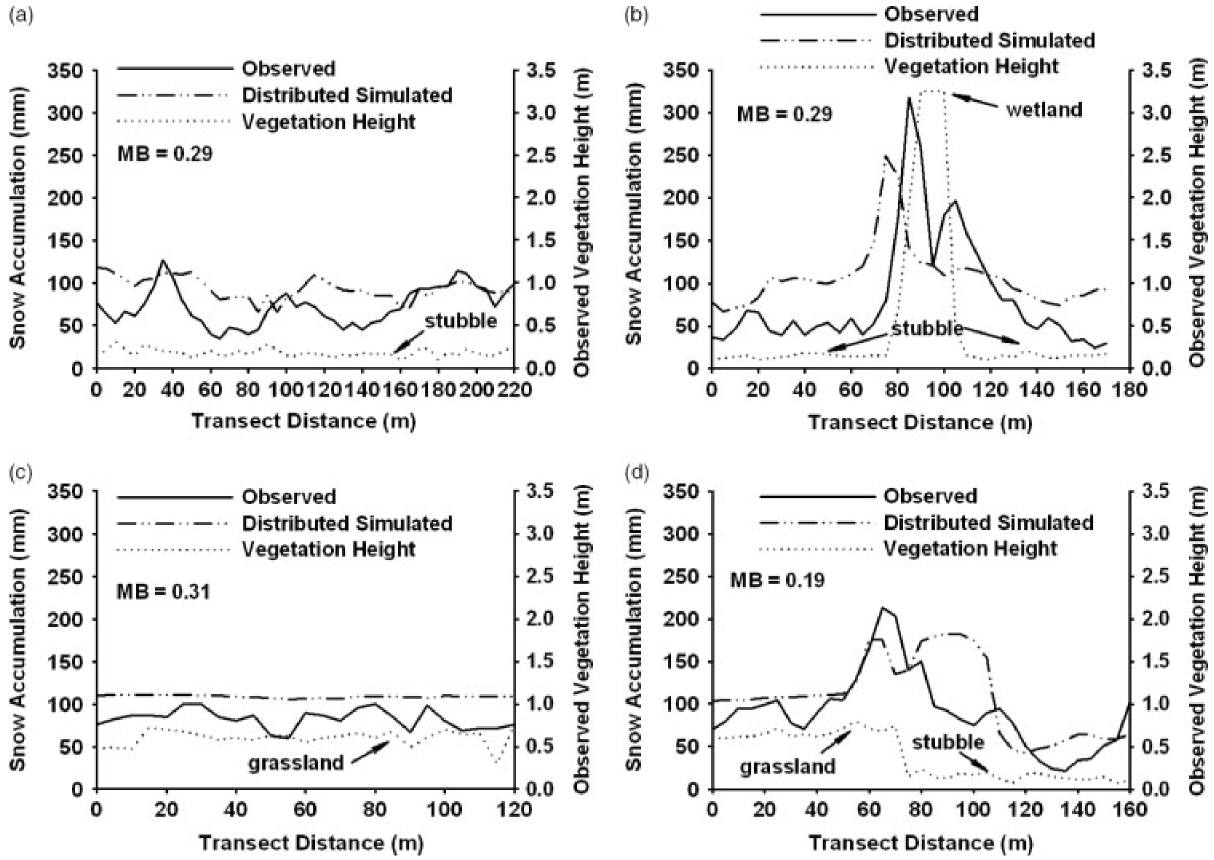

Figure 10. Comparison of the end-of-winter snow accumulation on 27 March 2006 between the spatially distributed approach and snow survey transects (a) transect 1, (b) transect 2, (c) transect 3 and (d) transect 4

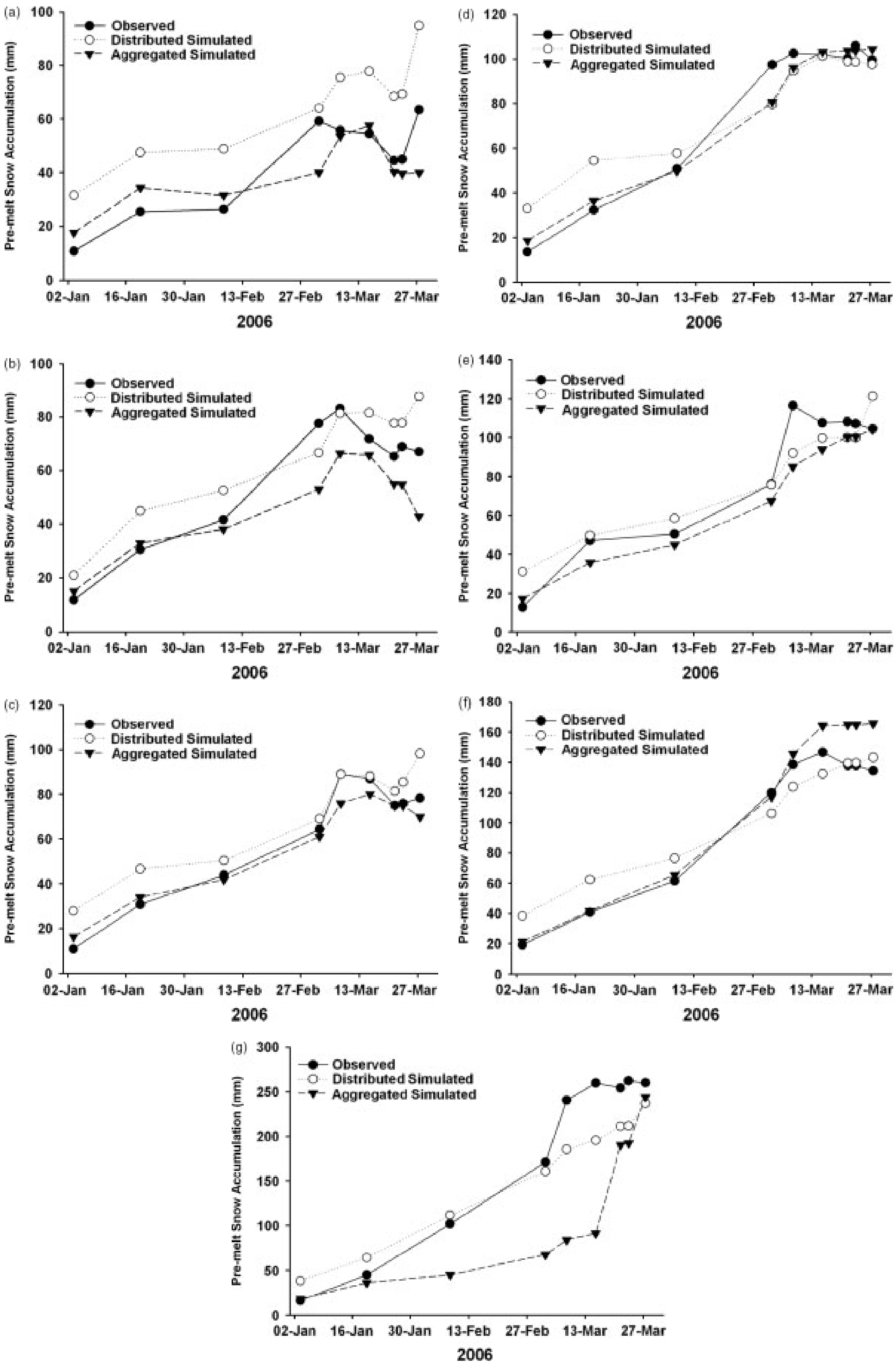

Figure 11. Comparison of the simulated pre-melt snow accumulation from the spatially distributed and spatially aggregated approaches to observed pre-melt snow accumulation on the HRUs (a) stubble hilltop, (b) stubble slope, (c) stubble level, (d) stubble valley, (e) grass level, (f) grass valley and (g) wetland

generate the end-of-winter snow mass moderately close to the observations on most HRUs; that is, the amount of overestimation or underestimation for total snow accumulation is within 30% as shown in Table III. However, for the 'stubble hilltop' and 'stubble slope' HRUs, the spatially distributed approach overestimated by more than 30%, whilst the spatially aggregated approach underestimated by more than 30%. This is likely due to uncertainty in the modelling of blowing snow over both complex topography and vegetation.

DISCUSSION

Figure 11(a)–(g) show that both the spatially distributed and spatially aggregated approaches had similar performances on most HRUs. The spatially distributed

DOI: 10.1002/hyp

approach predicted the development of snow accumulation better on some HRU (e.g. 'wetland'), while the spatially aggregated approach had better performance on other HRU (e.g. 'stubble hilltop'). However, the end-of-winter snow accumulation is the most important snow hydrology variable in the water balance.

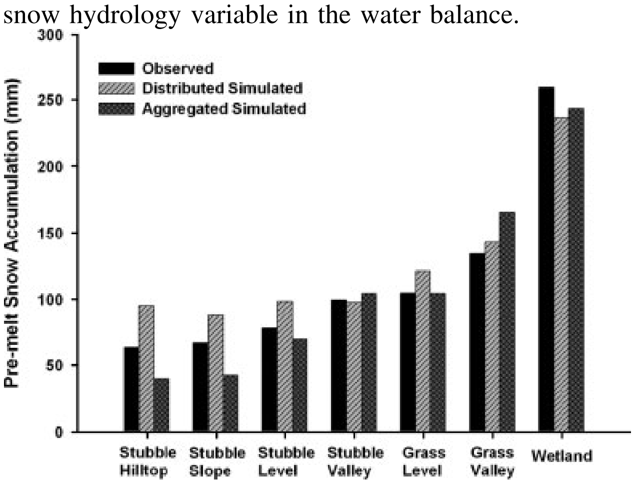

Figure 12. Comparison of the simulated cumulative pre-melt snow accumulation from the spatially distributed and spatially aggregated approaches to observed cumulative pre-melt snow accumulation on the HRUs

Figure 12 shows that both approaches provided similar predictability of the end-of-winter snow accumulation at the landscape scale when compared to the observations of cumulative winter snow accumulation. However, the computational time for the spatially distributed approach was about 30–50 min, while the spatially aggregated approach took about 1 min. Besides the time series snow accumulation shown in Figure 9, the spatially aggregated approach can also provide a visualization of spatial distribution of snow accumulation on different landscape units as illustrated in Figure 8. This visualization might not be as detailed as the one from the spatially distributed approach shown in Figure 7, but it is a useful portrayal of snow accumulation to HRUs and so suitable for further hydrological modelling. For the purpose of estimating the timing of pre-melt snow accumulation and total snow accumulation to landscape types in a small Canadian Prairie basin, the spatially aggregated approach was sufficiently accurate, relatively easy to parameterize and is computationally efficient. Landscape-scale snow accumulation information is required for interfacing blowing snow effects with land surface schemes or macroscale hydrological models where landscapes are represented as 'tiles' (Davison *et al.*, 2006; Dornes *et al.*, 2008). However, the spatially distributed approach provides much more detailed information on the location and spatial arrangement of deep snow drifts and scour zones that are of hydroecological and snow engineering interest.

Because discharge to and from wetlands is low due to the poor connectivity, blowing snow redistribution to wetlands from surrounding fields plays a critical role in Prairie hydrology by maintaining wetland water balances. Both spatially distributed and aggregated approaches were able to accurately simulate blowing snow redistribution to wetlands. Observations and these simulations show that wetlands have much higher end-of-winter snow accumulation (2 to 6 times) than surrounding cultivated fields and grasslands. Blowing snow redistribution to wetlands is very sensitive to land cover in the surrounding area. Fang and Pomeroy (2008) showed a land cover change scenario in which the vegetation cover surrounding a wetland was suppressed, which dramatically increased the amount of blowing snow transport to a small wetland.

This paper has shown that the spatially distributed and aggregated approaches can generate comparable results for blowing snow redistribution and cumulative winter snow accumulation in prairie pothole regions. This has important implications in modelling prairie wetland hydrology. Calculations based on fine-scale grid cells require high-quality DEMs such as those derived using LiDAR. LiDAR is not available to many places in the Canadian Prairies and is expensive, being $250–290 per km^2. Thus, the spatially aggregated approach that uses coarser scale basin topography and vegetation cover based on aerial photographs, topographic maps and some field inspection is far less resource intensive to implement and model.

CONCLUSIONS

Both spatially distributed and aggregated blowing snow simulations described similar end-of-winter SWE at the landscape unit scale and indicated substantial redistribution of blowing snow from exposed sparsely vegetated sites across topographic drainage divides to the densely vegetated prairie wetlands. Both simulations also agreed well with snow survey observations. While the distributed calculations provide a fascinating and detailed visual image of the interaction of complex landscapes and blowing snow redistribution and sublimation, it is clear that blowing snow transport and sublimation calculations can be successfully aggregated to the spatial scale of the major landscape units in this environment. This means that it should be possible to represent blowing snow redistribution successfully in the prairies using meso- and macroscale hydrological models and land surface schemes that incorporate blowing snow transport and sublimation processes.

ACKNOWLEDGEMENTS

The authors would like to thank Mr. Michael Solohub and Mr. Tom Brown, Centre for Hydrology for assistance in field work and programming. The portable distributed blowing snow modelling code provided by Dr. Richard Essery, University of Edinburgh is greatly appreciated. Funding was received from the Canada Research Chairs Programme and the Drought Research Initiative (DRI), a network funded by the Canadian Foundation for Climate and Atmosphere Sciences (CFCAS) and from an NSERC

Discovery Grant. Archived field observations at St. Denis National Wildlife Area, courtesy of Dr. Garth van der Kamp, Environment Canada and Dr. Bing Si, Department of Soil Science, University of Saskatchewan were crucial to this research and are much appreciated.

REFERENCES

Brown T, Pomeroy JW. 1989. A blowing snow particle detector. *Cold Regions Science and Technology* **16**: 167–174. DOI: 10·1016/0165-232X(89)90017-7.

Conly FM, van der Kamp G. 2001. Monitoring the hydrology of Canadian prairie wetlands to detect the effects of climate change and land use changes. *Environmental Monitoring and Assessment* **67**: 195–215.

Davison B, Pohl S, Dornes PF, Marsh P, Pietroniro A, MacKay M. 2006. Characterizing snowmelt variability in a land-surface-hydrologic model. *Atmosphere-Ocean* **44**: 271–287.

Dornes PF, Pomeroy JW, Pietroniro A, Verseghy DL. 2008. Effects of spatial aggregation of initial conditions and forcing data on modelling snowmelt using a land surface scheme. *Journal of Hydrometeorology* **9**: 789–803. DOI: 10·1175/2007JHM958·1.

Essery R, Li L, Pomeroy JW. 1999. A distributed model of blowing snow over complex terrain. *Hydrological Processes* **13**: 2423–2438.

Essery R, Pomeroy JW. 2004. Vegetation and topographic control of wind-blown snow distributions in distributed and aggregated simulations for an arctic tundra basin. *Journal of Hydrometeorology* **5**: 735–744.

Evans BM, Sheeder SA, Lehning DW. 2003. A spatial technique for estimating streambank erosion based on watershed characteristics. *Journal of Spatial Hydrology* **3**(1): 1–13.

Fang X, Pomeroy JW. 2008. Drought impacts on Canadian prairie wetland snow hydrology. *Hydrological Processes* **22**: 2858–2873. DOI: 10·1002/hyp.7074.

Godwin RB, Martin FRJ. 1975. Calculation of gross and effective drainage areas for the Prairie Provinces. *Proc. Canadian Hydrology Symposium*, Winnipeg, Manitoba, Associate Committee on Hydrology, National Research Council of Canada, 219–223.

Gray DM, Landine PG. 1988. An energy-budget snowmelt model for the Canadian Prairies. *Canadian Journal of Earth Sciences* **25**: 1292–1303.

Hayashi M, van der Kamp G, Rudolph DL. 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 1. Water balance. *Journal of Hydrology* **207**: 42–55.

Hayashi M, van der Kamp G, Schmidt R. 2003. Focused infiltration of snowmelt water in partially frozen soil under small depressions. *Journal of Hydrology* **270**: 214–229.

Jackson PS, Hunt JCR. 1975. Turbulent wind flow over a low hill. *Quarterly Journal of the Royal Meteorological Society* **101**: 929–955.

Lapen DR, Martz LW. 1996. An investigation of the spatial association between snow depth and topography in a Prairie agricultural landscape using digital terrain analysis. *Journal of Hydrology* **184**: 277–298.

Liston GE, Sturm M. 1998. A snow-transport model for complex terrain. *Journal of Glaciology* **44**: 498–516.

MacDonald J, Pomeroy JW. 2008. Gauge undercatch of two common snowfall gauges in a prairie environment. *Proceedings of the Eastern Snow Conference* **64**: 119–126.

MacDonald MK, Pomeroy JW, Pietroniro A. 2009. Parameterizing redistribution and sublimation of blowing snow for hydrological models: tests in a mountainous subarctic catchment. *Hydrological Processes* (in press). DOI: 10.1002/hyp.7356.

Mason PJ, Sykes RI. 1979. Flow over an isolated hill of moderate slope. *Quarterly Journal of the Royal Meteorological Society* **105**: 383–395.

Nash JE, Sutcliffe JV. 1970. River flow forecasting through conceptual models. Part I—A discussion of principles. *Journal of Hydrology* **10**: 282–290.

Pomeroy JW. 1989. A process-based model of snow drifting. *Annals of Glaciology* **13**: 237–240.

Pomeroy JW, Gray DM. 1995. Snowcover accumulation, relocation and management. National Hydrological Research Institute Science Report No. 7, Environment Canada, Saskatoon, 144.

Pomeroy JW, Li L. 2000. Prairie and Arctic areal snow cover mass balance using a blowing snow model. *Journal of Geophysical Research* **105**: 26619–26634.

Pomeroy JW, Gray DM, Landine PG. 1993. The prairie blowing snow models: characteristics, validation, operation. *Journal of Hydrology* **144**: 165–192.

Pomeroy JW, Marsh P, Gray DM. 1997. Application of a distributed blowing snow model to the arctic. *Hydrological Processes* **11**: 1450–1464.

Pomeroy JW, de Boer D, Martz LW. 2007a. Hydrology and water resources. In, *Saskatchewan: Geographic Perspectives*, Thraves B, Lewry ML, Dale JE, Schlichtmann H (eds.), CRRC: Regina, 63–80.

Pomeroy JW, Nicholaichuk W, Gray DM, McConkey B, Granger RJ, Landine PG. 1990. Snow management and meltwater enhancement final report. NHRI No. CS-90021, Report to the Saskatchewan Agriculture Development Fund, Environment Canada, Saskatoon, 164.

Pomeroy JW, Gray DM, Shook KR, Toth B, Essery RLH, Pietroniro A, Hedstrom N. 1998. An evaluation of snow accumulation and ablation processes for land surface modelling. *Hydrological Processes* **12**: 2339–2367.

Pomeroy JW, Gray DM, Brown T, Hedstrom NR, Quinton W, Granger RJ, Carey S. 2007b. The Cold Regions Hydrological Model, a platform for basing process representation and model structure on physical evidence. *Hydrological Processes* **21**: 2650–2667. DOI: 10·1002/hyp.6787.

Spence C. 2007. On the relation between dynamic storage and runoff: a discussion on thresholds, efficiency, and function. *Water Resources Research* **43**: W12416. DOI: 10·1029/2006WR005645.

Steppuhn H. 1981. Snow and agriculture. In *Handbook of Snow: principles, processes, management & use*, Gray DM, Male DH (eds.), Pergamon Press Canada Ltd., Ontario, 60–125.

Steppuhn H, Dyck GE. 1974. Estimating true basin snowcover. *Advanced Concepts and Techniques in the Study of Snow and Ice Resources*, U.S. National Academy of Sciences, Washington, D.C., 304–313.

Töyrä J. 2005. *Metadata for Preliminary LiDAR DEM—St. Denis NWA*. The National Water Research Institute, Environment Canada: Saskatoon, Canada, 5.

van der Kamp G, Hayashi M, Gallén D. 2003. Comparing the hydrology of grassed and cultivated catchments in the semi-arid Canadian prairies. *Hydrological Processes* **17**: 559–575. DOI: 10·1002/hyp.1157.

Walmsley JL, Taylor PA, Salmon JR. 1989. Simple guidelines for estimating windspeed Variations due to small-scale topographic features—an update. *Climatological Bulletin* **23**: 3–14.

Woo MK, Rowsell RD. 1993. Hydrology of a prairie slough. *Journal of Hydrology* **146**: 175–207.