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Memory effects of depressional storage in Northern Prairie hydrology

Kevin R. Shook* and John W. Pomeroy
Centre for Hydrology, University of Saskatchewan, Saskatoon, Canada, S7N 5C8

Abstract:

The hydrography of the Prairies of western Canada and the north-central United States is characterized by drainage into small depressions, forming wetlands rather than being connected to a large-scale drainage system. In droughts, many of these water bodies completely dry up, while in wet periods, their expansion can cause infrastructure damage. As wetlands expand and contract with changing water levels, connections among them are formed and broken. The change in hydrographic connectivity dynamically changes the hydrological response of basins by controlling the area of the basin which contributes discharge to local streams.The objective of this research was to determine the behaviour of prairie basins dominated by wetlands through two sets of simulations. The first consisted of application and removal of water (simulating runoff and evaporation) from a LiDAR digital elevation model (DEM) of a small basin in the south-east of the Canadian Province of Saskatchewan. Plots of water surface area and of contributing area against depressional storage showed evidence of hysteresis, in that filling and emptying curves followed differing paths, indicating the existence of memory of prior conditions. It was demonstrated that the processes of filling and emptying produced differing changes in the frequency distributions of wetland areas, resulting in the observed hysteresis. Because the first model was computationally intensive, a second model was built to test the use of simpler wetland representations. The second model used a set of interconnected wetlands, whose frequency distribution and connectivity were derived from the original LiDAR DEM. When subjected to simple applications and removal of simulated water, the second model displayed hysteresis loops similar to those of the first model. The implications for modelling prairie basins are discussed.

KEY WORDS prairie; wetland; memory; contributing area; hysteresis; connectivity

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INTRODUCTION

The Prairie region of western Canada and the northern United States is characterized by a cold, semi-arid climate and low relief. Consequently, much of the region does not possess a well-defined drainage system of the conventional type, and much of the land drains internally to small wetlands or sloughs (Stichling and Blackwell, 1957). Water in the wetlands will evaporate during the open-water season and will also contribute to evapotranspiration of riparian vegetation. Some water may infiltrate below the bottom of the wetland, providing recharge to the local groundwater. In many Prairie locations, the very low hydraulic conductivities of the clayey tills underlying surface deposits can effectively restrict deep percolation (Woo and Rowsell, 1993). The presence of aquitards, along with the semi-arid nature of the prairie climate, causes many prairie streams to exhibit little in the way of baseflow, despite the storage of water over multiyear periods in surface depressions.

Because of the internal surface drainage, and general lack of subsurface flow, much of the region is designated as being non-contributing to streamflows (Godwin and Martin, 1975). The extent of the non-contributing region, for streams which drain into Canadian rivers, is plotted from data obtained from Agriculture and Agri-Food Canada (AAFC) (http://www.rural-gc.agr.ca/pfra/gis/gwshed_e.htm) in Figure 1. The AAFC non-contributing regions are identified as those which do not contribute to downstream flow for a median (i.e. 1:2 year) annual runoff. However, previous research has shown that the extent of the non-contributing fraction of any basin region is dynamic and changes with the amount of water in depressional storage (Stichling and Blackwell, 1957; Pomeroy *et al*., 2010). The dynamic contributing area of depressional storage has been compared to the Variable Source Area concept (Bowling *et al*., 2003; Spence, 2010).

Other important characteristics of the wetland depressional storage include:

1. the connectivity between wetlands is ephemeral, occurring only when wetlands are full,
2. the magnitudes of the depth, area and volume of water stored in wetlands vary tremendously as the wetlands range from transient puddles to semi-permanent lakes, and
3. the area of a wetland exposed to evaporation is a non-linear function of its depth (Hayashi and van der Kamp, 2000).

*Correspondence to: Kevin R. Shook, Centre for Hydrology, University of Saskatchewan, Saskatoon, Canada, S7N 5C8.
E-mail: kevin.shook@usask.ca

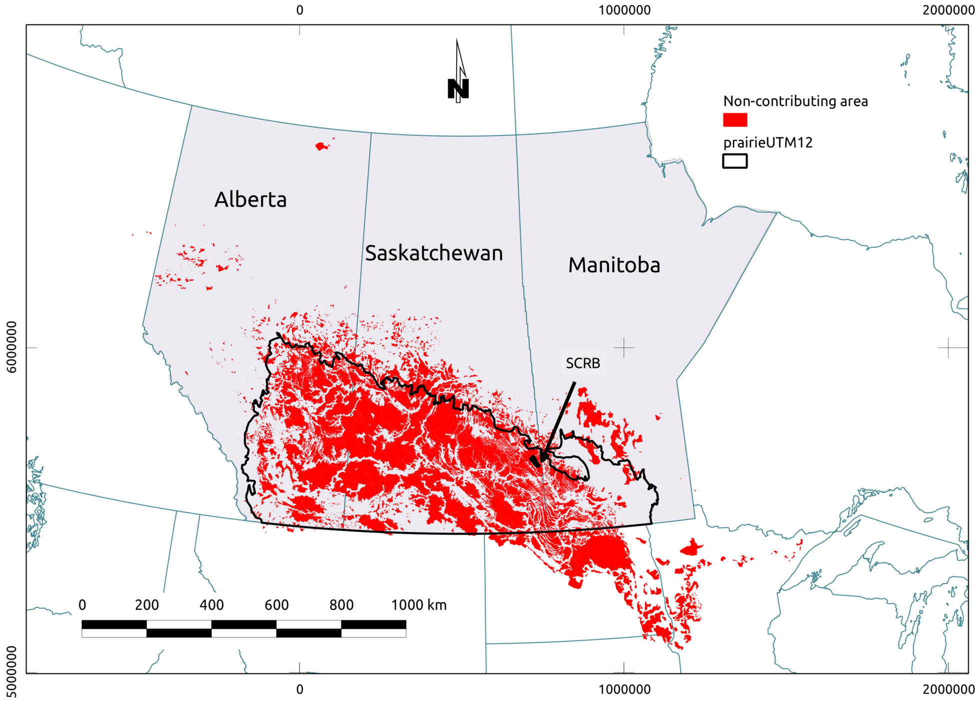

Figure 1. Extent of non-contributing regions in the Canadian Prairies showing the location of the Smith Creek Research Basin. The projection used is UTM 13

These characteristics have been shown (Spence, 2007; Shaw, 2010) to contribute to nonlinearity in the relationship between the fraction of a basin contributing runoff and the water storage in the basin. In addition, the depressional storage of water may cause 'memory'in the system, where the response of the system at any time depends on the history of inputs and outputs. Shaw (2010) established some theoretical relationships for wetland filling in simple wetland systems; however, the question of the response of wetland complexes (i.e. of large numbers of interconnected wetlands) to the hydrological inputs and withdrawals has not been addressed.

Prairie wetland hydrology is extremely complex (Woo and Rowsell, 1993; Hayashi *et al*., 1998; van der Kamp and Hayashi, 2003; Fang *et al*., 2010; Pomeroy *et al*., 2010), and it is necessary to reproduce the sequence of hydrological events to understand the response of wetland complexes to forcings. Prior research has been limited to considering fewer than ten wetlands (Shaw, 2010), while even small prairie basins may have many thousands of wetlands (Zhang *et al*., 2009).

Hysteresis has been observed in many hydrological relationships, from soil pore to basin scales (Prowse, 1984; O'Kane, 2005; O'Kane and Flynn, 2007; Spence, 2010). The Preisach model of hysteresis is widely used and has been applied to hydrological systems (O'Kane and Flynn, 2007). The Preisach model requires that the hysteresis loops exhibit return-point memory (all minor loops return to their starting point) and that parallel loops are congruent (Flynn *et al*., 2006).

O'Kane (2006) demonstrated that collections of linear reservoirs can exhibit hysteresis. However, the storage and fill and spill behaviour of water in prairie wetlands is very different from that of a conventional reservoir whose discharge is proportional to the depth of water in the reservoir. As described above, due to the presence of aquitards, the contribution of depressional storage to discharge through groundwater flow is typically negligible. The water in depressional storage never directly contributes to streamflow. As wetlands fill, their water surface areas increase, and small wetlands will combine to form larger wetlands. When enough wetlands connect, they will form a path to the stream. At this point, the wetlands are fully connected, and any further additions of water will contribute flow to the stream. However, the water below the sill elevations does not contribute flow – it is dead storage. The memory effect of depressional storage is not due to the dead storage, except in that the dead storage sets the contributing area.

The overall objective of this research is to quantify the causes of the nonlinearities and memory effects caused by depressional storage in prairie basins. A secondary objective is to determine the extent to which models of wetland complexes must reproduce the number of wetlands and their arrangements, to produce realistic responses to inputs, to allow more accurate modelling of wetland-dominated basins.

METHODOLOGY

This research is comprised of model simulations of the responses of prairie wetlands to forcings of runoff and evaporations. In the context of this research, the term 'model' is used to refer to the method of simulating the behaviour of the wetlands, rather than the computer program used for the simulation.

Two models were used to determine the response of sub-basin 5 to the addition and removal of water, but both only attempted to simulate the surface responses of the sub-basin; no attempt was made to reproduce groundwater contributions to streamflow, which are very small in the research basin and can be neglected (Fang *et al*., 2010). Model 1 was intended to determine the spatial distribution of surface runoff as a function of basin state, by the addition and removal of specified quantities of water. It was also intended to test for the existence of nonlinearity and/or hysteresis in the responses of the basin. As described below, Model 1 was complex model of the basin's topography, which was tested with very simplified fluxes of water. Model 2 was intended to determine the ability of a simpler model to reproduce the types of behaviours shown by Model 1. Model 2 was a conceptual model based on statistics drawn from the complex topography of Model 1, which was tested using the same types of simplified fluxes as applied to Model 1.

Research location

The Smith Creek Research Basin (SCRB) in south-eastern Saskatchewan, shown in Figure 1, was selected for this research. The basin is flat with elevations ranging between 490 and 548 m, and with slopes between 2% and 5%. Land use is dominated by pasture cropland with the primary crops being cereal grains and oilseeds. There are patches of deciduous woodland, particularly near to wetland locations. The basin was selected because it is dominated by wetlands, over 10,000 wetlands larger than 100 m² having been identified (Fang *et al*., 2010), and the basin possesses an ephemeral stream. The area of SCRB is 445 km² and has been divided into five sub-basins, as shown in Figure 2. This research was conducted exclusively using data for sub-basin 5, which is approximately 12 km² in area.

Model 1, LiDAR DEM-based runoff simulation

Model 1directly simulated surface runoff from a LiDAR digital elevation model (DEM) of SCRB sub-basin 5. The forcings applied to Model 1 were very simplified to minimize computational effort. Arbitrary depths of water were applied and removed, as precipitation and evaporation. The applied precipitation was spatially uniform. Only direct evaporation from the free-water surface was computed, and this was also assumed to be spatially uniform. The spatial uniformity of the fluxes implied that that any nonlinearity in response of the basin was due to the properties of the DEM.

Computational effort was also reduced by ignoring the rate of runoff. No time step was used. In each simulation run, the runoff water was allowed to flow over the DEM until the system reached a steady state. The state of the simulated basin after each fully drained simulation was analogous to the state of a natural basin after a heavy rainstorm or snow melt event, when surface runoff had ceased. The software which computed the flows was comprised of three custom-written Fortran 95 programs, 'RUNOFF', 'DRAIN' and 'EVAP'.

The program 'RUNOFF' applied a specified depth of simulated water to the DEM and allowed the water to run from the relatively higher locations on the DEM and accumulate in depressional storage. As this program did not allow any water to leave the basin, the edge of the DEM acted as a dam at the mouth of the creek and artificially caused runoff water to backup from the creek over the DEM. The non-creek portion of the DEM did not contact the edge of the DEM, as it was surrounded by a 'mask' of cells which marked the basin divide.

The program 'DRAIN' allowed water to exit the DEM. This program used the same algorithm to redistribute water as did 'RUNOFF', the only difference being that 'DRAIN' allowed water to exit from the minimum-elevation cell, which was located within Smith Creek.

The program 'EVAP' simulated evaporation from the water on the DEM. No attempt was made to compute evaporative fluxes; a specified depth of water (limited to the existing depth of water) was removed from each cell.

Using the three programs, any desired sequence of filling or evaporation events could be simulated over the DEM. The water distribution algorithm used by 'RUNOFF' and 'DRAIN' is based on that of Shapiro and Westervelt (1992). The algorithm was selected because it explicitly includes the effects of depressions. Other methods, such as the well-known D8 algorithm, generally require the user to fill depressional storage before determining a single flow path for each cell (Garbrecht and Martz, 1997). When runoff is computed by the Shapiro and Westervelt algorithm, depressions trap water until they are filled to their sill elevations, after which the addition of further water causes runoff downslope(s). Thus, the simulated basin will respond dynamically to the addition and removal of water in much the same way as areal basin.

The algorithm consists of the following steps:

1. For each iteration, each cell in the DEM is selected in turn, and its water surface elevation is found from the sum of the cell's DEM elevation and its water depth.
2. The water surface elevation is found for each of eight the cells immediately adjacent to the cell of interest.
3. The difference between the water surface elevation of the cell of interest and of each its neighbours is computed.
4. Where there is a difference in water surface elevation, a volume of water equivalent to one eighth of the difference in surface elevation multiplied by the cell area is transferred to the cell having the lower water surface elevation.

The Shapiro and Westervelt algorithm is iterative, as the distribution of water is based on the relative water depth of neighbouring cells, which is temporally variable. As water is added only in the first iteration, the spatial arrangement of water on the DEM will eventually converge to a final distribution. In practice, convergence is deemed to occur when the error (the maximum change in water depth, determined every 100 iterations), is smaller than a pre-defined tolerance.

The sub-basin was characterized using a DEM derived from LiDAR data (Fang *et al*., 2010; Pomeroy *et al*., 2010).

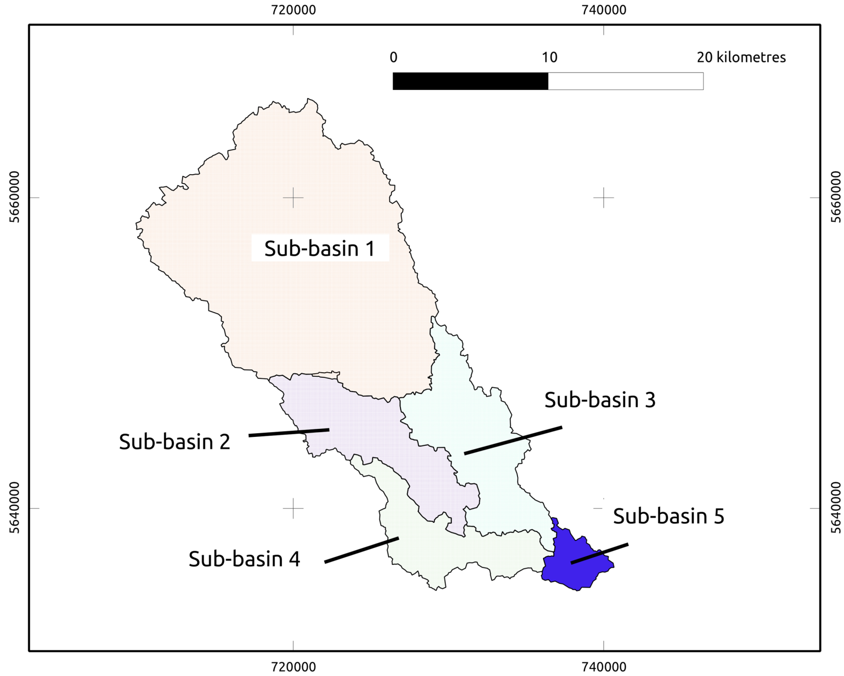

Figure 2. Arrangement of sub-basins in Smith Creek Research Basin. The projection used is UTM 13

The original LiDAR data were collected at a horizontal resolution of 1 m during the interval of October 14 to 16, 2008. The collection procedures of the LiDAR data are described in detail by LiDAR Services International (2009), indicating elevation RMS errors of 0.05 m at SCRB. To reduce the quantity of data to a manageable size, the horizontal resolution of the LiDAR elevations was reduced to 10 m by resampling (Fang *et al*., 2010). Because some water was present in the depressions when the LiDAR data were collected, all modeling was done relative to this initial state.

The number of iterations required for convergence of the Shapiro and Westervelt algorithm depended on the size of the DEM array, the depth of water added and the tolerance selected. On SCRB sub-basin 5 at a spatial resolution of 10 m, additions of several hundred millimetres of simulated excess precipitation required as many as 200,000 iterations to converge to a tolerance of 1 mm. Because a single execution of the programs 'RUNOFF' or 'DRAIN' could take as long as 12 h, and because tens of runs were required to demonstrate the dynamics of the system, computing facilities were obtained from Compute Canada to allow many simultaneous model runs.

Model 2, a conceptual model of wetland hydrography

Because of its simplified processes and high computational cost, Model 1 is not suitable for modelling the temporal responses of prairie basins. Shaw (2010) used a simple conceptual model of small numbers (less than 10) of individual wetlands. Each wetland was modelled as a simple reservoir, which spilled to a neighbouring wetland when the water level exceeded the sill elevation. The combination of wetlands responded in complex ways, when subjected to inputs of simulated rainfall, runoff and evaporation.

The very large numbers of wetlands in prairie basins make the modelling of each wetland impracticable. Zhang *et al.* (2009) found that the frequencies of prairie wetland and lake areas could be described by power-law relationships. As prairie wetlands in a given region can be treated as being members of a frequency distribution, a conceptual model of an areal fraction of a prairie basin can be assumed to be statistically representative of the entire basin, if the model's wetlands are also representative of the frequency distribution of wetlands in the entire basin.

The memory effects of prairie wetlands can be modelled by forcing the conceptual model of wetlands with simulations of prairie hydrological processes. The conceptual model's dependency on model state should be similar to, and can be tested against the results of, the DEM model.

Connectivity of model wetlands. Shaw (2010) demonstrated that the connectivity of modelled wetlands can influence their collective responses to simulated inputs. Phillips *et al.* (2011) analysed lakes in a low-relief bedrock basin, showing that downstream lakes act as 'gatekeepers' controlling the contribution of upstream lakes to basin discharge. Because these lakes are similar in their thresholding behaviour to prairie wetlands, it is argued that the Model 2 wetlands must be connected in a similar manner to those of the actual basin at SCRB to produce valid simulations. As Model 2 used a statistically representative set of wetlands, their connectivity was derived from the statistical connectivity of the wetlands in SCRB.

The connectivity of the wetlands in SCRB was determined by delineating a dendritic drainage network on which were overlaid polygons of the wetlands network using QGIS (http://www.qgis.org). The drainage network was computed using the program TOPAZ (Fang *et al*., 2010) on a low-resolution (50 m) DEM resampled from the original high-resolution LiDAR DEM (Fang *et al*., 2010). The use of low-resolution data was dictated by the memory limitations and processing time requirements of TOPAZ. Although the TOPAZ data are at a lower resolution than that of the DEM used for runoff simulations, it was assumed that the low-resolution LiDAR data adequately represented the connectivity among wetlands in the basin. As is described below, the wetland connectivities determined are statistical ratios. Although use of higher resolution DEM data would certainly result in larger numbers of wetlands intersecting the drainage network, there is little reason to believe that this would change the ratios used to establish the statistical connectivities.

The wetland connectivity was based on Horton's law of stream numbers (Horton, 1945) given by

$$N_o = r_b^{s-o}, \tag{1}$$

where

N_o = number of streams of order o,
r_b = bifurcation ratio, and
s = number of streams of highest order in the watershed.

The bifurcation ratio was defined as

$$r_b = \frac{N_o}{N_{o+1}}, \tag{2}$$

where

N_{o+1} = number of stream segments of next order.

The total number of wetlands intersecting each stream segment was tabulated by Horton-Strahler order and is shown in Table 1 with the associated bifurcation ratios. The bifurcation ratios are assumed to represent the connections between wetlands, indicating that each wetland is connected to between one and three wetlands of the next order. Shaw (2010) showed that the use of single strings of wetlands (i.e. bifurcation ratio =1) in a simulation can exaggerate the influence of a single wetland on simulated discharges, due to each wetland acting as a 'gatekeeper'. To avoid this situation, Model 2 only simulated Horton-Strahler orders of 1 through 5, thereby avoiding the additional single wetland for order 6. As Model 2 used a small number of simulated wetlands, integer bifurcation ratios of 2, 2, 2, 3 and 1 were used for Horton-Strahler orders of 1 through 5, respectively. The resulting conceptual wetland model consisted of upland and 46 wetlands, whose arrangement is shown schematically in Figure 3.

Model 2 program algorithm. Model 2 was executed by a purpose-written Fortran 95 program SIMPLE, named for its simplified forcings which were similar to those of Model 1. Like Model 1, Model 2 does not have a time step, but simulates the response of wetlands to the addition and removal of discrete quantities of water.

Table I. Total number of Smith Creek Research Basin wetlands and bifurcation ratios for each Horton-Strahler stream order

Horton-Strahler stream order	Wetland count	Bifurcation ratio
1	2837	1.96
2	1445	1.66
3	868	2.24
4	388	2.69
5	144	0.99
6	145	

The addition of water is simulated by the following steps.

1. A uniform depth of water is added to all wetlands. The depth applied is increased by the ratio of total basin area to total maximum wetland area, i.e. 4, to allow for runoff from the area outside the wetlands.
2. Evaluating from furthest upstream to the outlet, each wetland is tested to see if the total volume of water stored is greater than the maximum permitted. If so, the difference is routed downstream.
3. Because wetlands are connected in parallel as well as in series, the water redistribution is iterated until the maximum quantity of water moved is smaller than a pre-set tolerance. The tolerance used, 0.1 m³, is equivalent to 1 mm over the smallest wetland, as used by Model 1.

SIMPLE simulates evaporation by the removal of a uniform depth of water from each wetland which requires modelling the relationships among depth (which determines the exceedence of the sill elevation), water surface area (which affects evaporation) and volume.

As water is added to and removed from wetlands, the area of the water surface changes. Hayashi and van der Kamp (2000) showed that the relations between water surface area (*A*) and volume (*V*) and water depth (*h*) typically vary according to

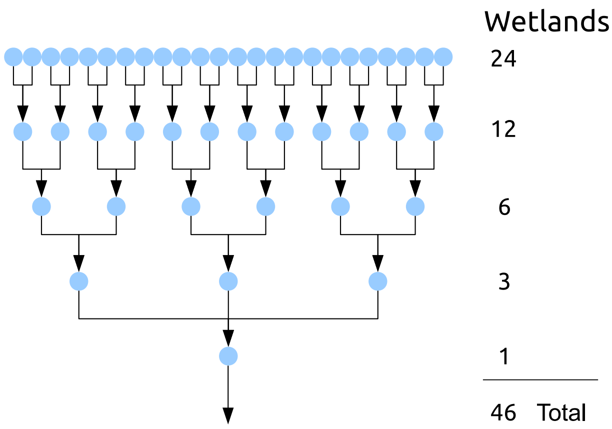

Figure 3. Schematic arrangement of simulated wetlands of Model 2

$$A = s\left(\frac{h}{h_o}\right)^{2/p}, \textbf{\textit{and}} \tag{3}$$

$$V = \frac{s}{1+2/p}\left(\frac{h^{1+2\left(\frac{2}{p}\right)}}{h_o^{2/p}}\right), \tag{4}$$

where

h_o = unit depth (1 m), and
s, p = constants.

Minke *et al.* (2010) developed simplified methods for estimating s and p. Fang *et al.* (2010) showed that these methods could be used to estimate mean values for a region from LiDAR DEM data, finding p =1.72 for wetlands whose maximum area was smaller than 10,000 m², and $p = 3.3$ for larger wetlands, at SCRB.

The wetlands used in Model 2 were selected randomly from the set of SCRB wetlands, using the maximum values of A and V determined by Fang *et al.* (2010). Using the Fang *et al.* (2010) values for p, Equations 3 and 4 were solved for s, using $h = h_{max}$, for each wetland. Having computed s, A and V were calculated for any wetland for values of h smaller than h_{max}.

SIMULATION RESULTS

The results of the simulations are presented separately to demonstrate their similarities and differences.

Model 1 simulation results

Examples of the final water spatial distributions produced by Model 1 are plotted in Figure 4 for additions of 25 and 100 mm, and the subsequent removal of 100 mm, of water. As expected, the simulations resulted in water accumulation in discrete wetlands. Increasing the quantities of water added increased the sizes of the wetlands, with smaller wetlands frequently joining to form larger wetlands. The removal of 100 mm of water did not return the water accumulation to its original state, as the water was applied to the entire DEM, but was only removed from the wetlands. The effects of water addition and removal on the frequency distribution of wetland areas are addressed below.

Figure 5 plots the relationship between the fractional water-covered area and the fractional water storage of the simulated basin for several simulations. The fractional water-covered area is the area of the basin covered by water divided by the maximum possible water-covered area. The fractional water storage is the volume of water contained in depressional storage divided by the maximum possible depressional storage. The fractional water-covered area is of interest as a diagnostic variable of the basin's state because it is potentially measureable by remote sensing.

The plotted points describe obvious hysteresis loops. Loop 1 was caused by simple wetting (cumulatively adding excess precipitation until filled) and drying (cumulatively evaporating water). Approximately 300 mm of water was required to completely fill all of the storage in the DEM. The lower (drying) limb of Loop 2 resulted from removing up to 100 mm from an initial addition of 100 mm. The upper (wetting) limb resulted from the addition of up to another 100 mm of additional water. The lower (drying) limb of Loop resulted from removing up to 200 mm from an initial addition of 100 mm. The upper (wetting) limb of the loop resulted from the addition of up to 50 mm.

Figure 6 plots the fractional contributing area (the fraction of the basin contributing to runoff) against the fractional water storage of the simulated basin for the same sequences of adding and removing water depicted in Figure 5. The fractional contributing area was determined by adding an incremental depth of water (1 to 5 mm) to the DEM after each of the simulation runs used in Figure 5 and calculating the fraction of water running off.

The plots of fractional contributing area against the fractional water storage also show hysteresis, although the shapes of the three loops plotted are very different from those in Figure 5. In Loop 1, the fractional contributing area increases nearly linearly with the fractional volume until the entire basin contributes to discharge. The removal of water causes an abrupt decrease in contributing area to zero. Loops 2 and 3 shows a similar pattern, with drying causing a rapid decrease in contributing area to zero, followed by a rapid rise in contributing area with re-wetting.

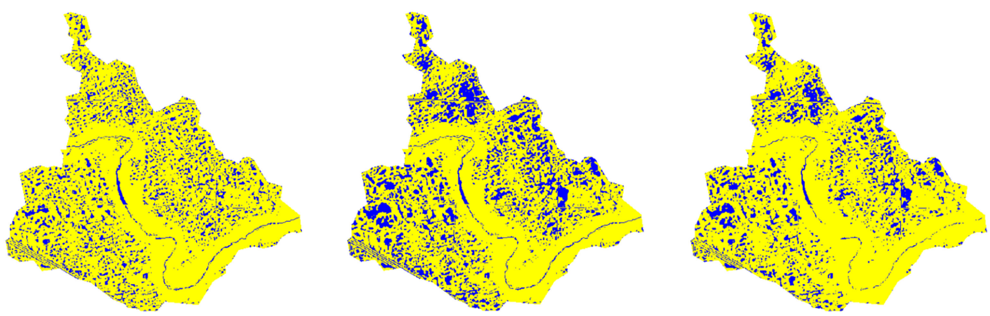

Figure 4. Spatial distribution of simulated water after three sequential runs of Model 1 for Smith Creek Research Basin sub-basin 5. The yellow regions are dry, the blue regions contain water. From left to right, the images represent the addition of 25 mm of water, the addition of a further 75 mm of water (total 100 mm) and the subsequent removal of 100 mm of water

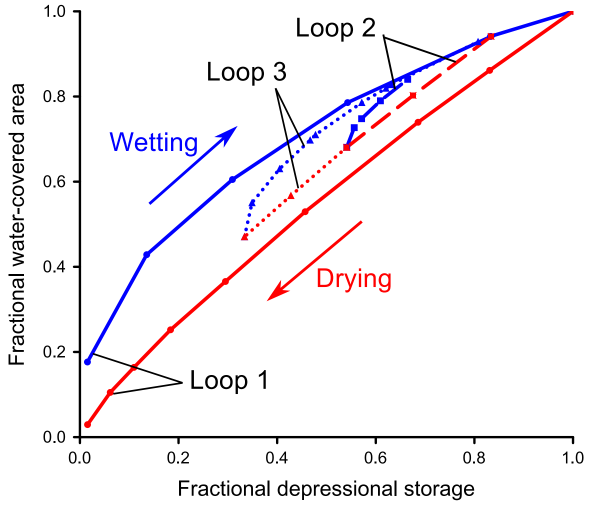

Figure 5. Modelled fractional water-covered area *versus* fractional water storage for Model 1 simulations. Loop 1 is caused by adding water until the DEM is filled, followed by removing water until the DEM is nearly empty. Loop 2 is caused by the addition of100 mm, followed by the removal of 100 mm, and re-filling. Loop 3 is caused by the addition of 100 mm, the removal of 200 mm, and re-filling

The simulated hysteresis between water coverage and depressional capacity and between contributing area and depressional capacity are potentially important to prairie hydrology. The dependence of the contributing area on the state of wetland storage prevents the use of hydrological models which assume a constant contributing area. Furthermore, the hysteresis and nonlinearity in the relationship between the areal water coverage and the storage prevent the use of a simple relationship to relate the contributing area to the depressional storage.

In Figure 5, the filling and emptying loops between water-covered area and storage show return-point memory in that they return to their original starting points. The loops between contributing area and storage in Figure 6 do not appear to show return-point memory. The shapes of the Model 1 hysteresis loops are unlike any of the typical hysteresis loops characterized by Lapshin (1995), many of which are sigmoidal. As the Model 1 envelope curves are not sigmoidal in shape, the distance between the curves is not consistent. Thus, parallel loops within the envelope curves will not be congruent, as their lengths will differ, and therefore the Preisach model is not a good descriptor of the hysteresis associated with prairie wetlands.

Wetland area frequency distribution. The areas of wetlands resulting from the Model 1 simulations were determined using the module 'r.le' of the open source GIS program GRASS (Baker and Cai, 1992). In Figure 7, the exceedence frequencies plotted against wetland areas clearly demonstrate that the frequency distributions of the wetland areas exceeding 1000 m^2 are well-described by Pareto (power-law) distributions. Similar relationships were also found by Zhang *et al.* (2009) for remotely sensed measurements of real wetlands.

The application and removal of water affect the frequency distributions of open water areas in differing ways, as was also found by Zhang *et al.* (2009). Increasing the amount of water added from 25 to 100 mm appears to rotate the frequency distribution counter-clockwise, shifting the curve upward and to the right. Conversely, removal of water shifts the distribution upward and leftward. The differing trajectories of the distribution show that the frequency distribution plot of the open water area will follow a looping pattern as precipitation and evaporation alternate.

The differing frequency distribution trajectories explain the observed hysteresis loops in the filling and emptying curves of fractional water-covered area and contributing area. The wetland frequency distribution at any time depends on the prior history of filling and emptying events. Therefore, the total open-water and the contributing area,

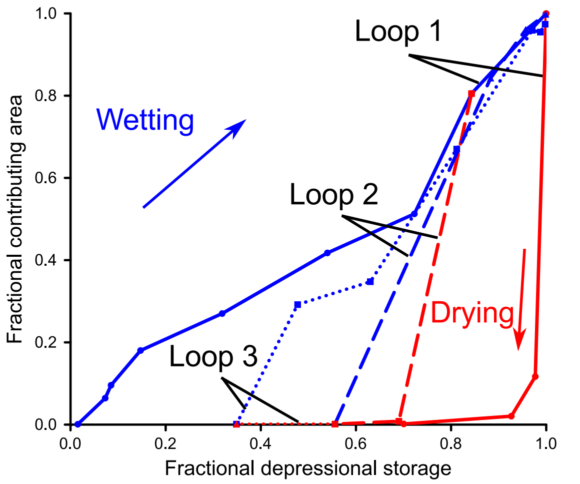

Figure 6. Modelled fractional contributing area *versus* fractional water storage for Model 1 simulations. Loops 1, 2 and 3 have the same additions and removals of water as in Figure 5

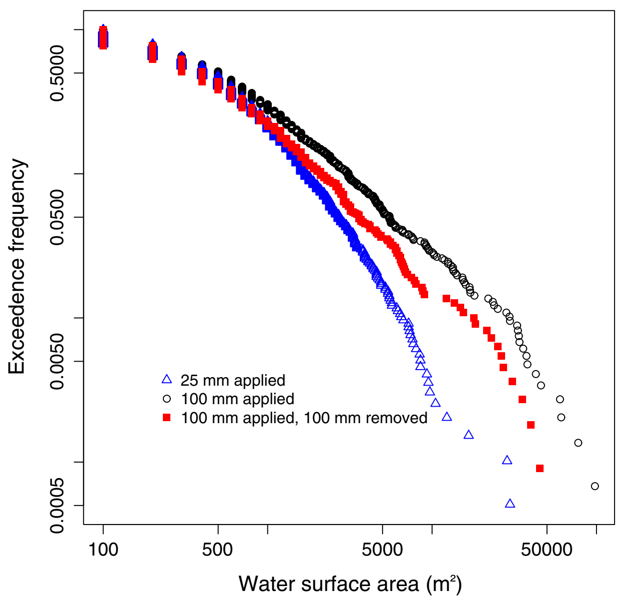

Figure 7. Frequency distributions of water surface areas resulting from three simulation runs using Model 1. The scales are logarithmic. The plots represent the addition of 25 mm of water, the addition of a further 75 mm of water (total 100 mm) and the subsequent removal of 100 mm of water

which is an index of wetland connectivity, will also depend on the history of events.

Model 2 Simulation results

Because the model uses a small number of simulated wetlands, its behaviour is affected by the particular set of wetlands selected. The degree to which this is true was tested by running the program using (a) a single set of 46 wetlands and (b) eight sets of 46 wetlands (i.e. 368 wetlands).

The water area *versus* volume curves are plotted in Figure 8, along with the envelope curves for Model 1, for curves of simple addition and removal of water. As expected, the Model 2 curves lie within the Model 1 curves, indicating that the degrees of hysteresis produced by the two models are very similar. As was also expected, the simulation using eight sets of wetlands produced curves more similar to those of Model 1 than did simulations using one set of wetlands.

The contributing area *versus* volume curves of Model 2 are plotted in Figure 9, along with the envelope curves corresponding to Model 1. The curve corresponding to a single set of wetlands resembles a staircase with horizontal ranges which correspond to the filling of large 'gatekeeper' wetlands. Using eight sets of wetlands in parallel reduced the influence of any single large wetland and resulted in a smoother volume–area hysteresis curves which was more similar to those of Model 1. As do the Model 1 curves, the Model 2 curves show the contributing area increasing as water is added, and dropping rapidly to zero as water is removed. Unexpectedly, both runs of Model 2 produced much smaller fractional contributing areas than did Model 1, when the magnitude of the fractional depressional storage was smaller than approximately 0.7.

The Model 1 curve indicates that a portion of the basin contributes to discharge almost immediately upon the addition of water, while the Model 2 curves show no contributing area until the fractional depressional storage is approximately 21%, which corresponds to the filling of the

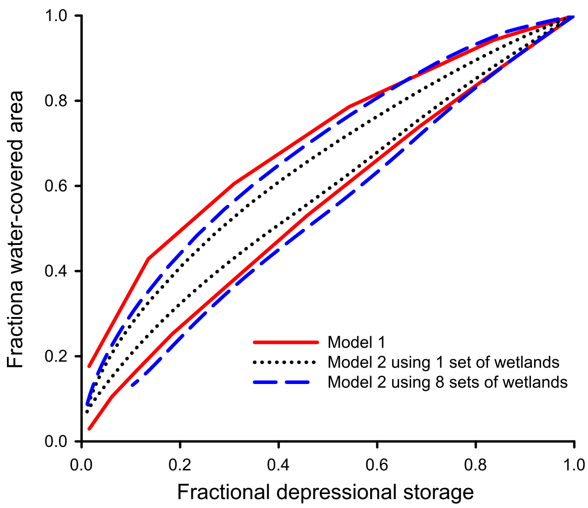

Figure 8. Envelope curves of fractional water-covered area *versus* fractional water storage for Model 1 and Model 2 using one and eight sets of 46 simulated wetlands

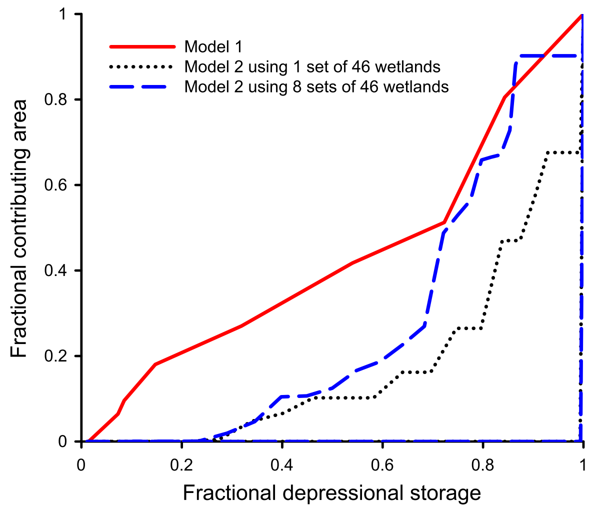

Figure 9. Envelope curves of fractional contributing area *versus* fractional water storage for Model 1 and Model 2 using one and eight sets of 46 simulated wetlands

smallest wetlands of Fang *et al.* (2010), which had maximum storages of approximately 200 mm. Evidently, the difference among the outputs of the models is due to differences in the frequency distributions of shallow wetlands. The cause is believed to be the omission of ephemeral wetlands, which were not detected by Fang *et al.* (2010).

SUMMARY AND CONCLUSIONS

A model of thousands of wetlands showed the existence of hysteretic relationships between the fractional water-covered area, the fractional contributing area of a basin and the fractional water storage. The cause of the observed hystereses appears to be the differences in the changes in the frequency distributions of the wetland areas due to runoff and evaporation processes. Evaporation is taken from all water surfaces equally, whilst runoff is delivered along topographic flow pathways to each wetland.

A simple model of a few wetlands can reproduce the types of hysteresis demonstrated by the DEM-based model of thousands of wetlands and show that wetland complexes 'remember' their initial conditions. The time scale(s) of wetland 'memory' may be estimated by forcing Model 1 or 2 with physically based fluxes, but will require either solving the computational requirements of Model 1, or determining the frequency distribution of ephemeral wetlands for Model 2.

The simulations presented have not yet been validated with field observations. Validation from field data will require measurements of input (runoff) and output (evaporation) fluxes, as well as measurements of the system outputs (streamflows) and state variables (depressional storages). Although all of these components have been estimated or measured at locations in the Canadian Prairies, they have rarely been measured at one location over an extended period of time.

The existence of hysteresis in the simulations indicates that models using single linear transfer functions cannot accurately estimate the dynamic contributing area of those prairie basins strongly affected by the depressional storage. The existence of hysteresis between the volume and area of water in depressional storage means that bulk estimates of total water surface area cannot be used as an index of water storage or contributing area.

Runoff events occur rarely in the prairies, generally during the spring snowmelt period (Gray, 1973), and the effects of wetland storages are only observed at this time. Each wetland's storage is a separate state variable, the magnitude of which changes throughout the open water season due to rainfall and evaporation. Therefore, it is nearly impossible to build a pure input–output model of a prairie basin dominated by wetlands as the magnitudes of the state variables change between observations, and the number of observations is very small.

Remote-sensing techniques, as used by Zhang *et al.* (2009) and Touzi *et al.* (2007), if capable of discriminating water from soil at scales sufficient to delineate individual wetlands, may provide a way of initializing wetland models. When forced by physically based models of fluxes, a statistically representative collection of model wetlands may be able to model the contributing area, and therefore the runoff, from a wetland-dominated prairie basin.

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