FORECASTING STREAMFLOW RUNOFF FROM SNOWMELT IN A PRAIRIE ENVIRONMENT

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ABSTRACT

The paper focusses on the effects of infiltration to frozen soils and the "contributing" area of a watershed on synthesizing streamflow hydrographs and estimating runoff volumes from snowmelt in a prairie environment. A "first-generation" model describing infiltration to frozen soils is presented and methods of interfacing it in an operational streamflow forecasting system discussed. The model divides frozen soils into three broad classes according to their infiltration potential: Unlimited, Limited and Restricted, based on the physical characteristics of the soils and surface conditions at the time of snowmelt. For soils of "Limited" potential, snowmelt infiltration is calculated from the snowcover water equivalent and the moisture (ice) content of the soil layer, 0-300 mm. It is demonstrated that the infiltration model significantly improves the performance of the National Weather Service River Forecasting System in simulating streamflow from snowmelt on the Wascana Creek basin at Sedley, Saskatchewan, which is located approximately 50 km southeast of Regina.

The interaction between runoff volume, unit runoff potential (the sum of the snowcover water equivalent plus precipitation during the melt period less the snowmelt infiltration potential) and the "apparent" area of the watershed contributing to streamflow is reviewed. A relationship between these variables is presented which is based on the assumption that the runoff release and response characteristics of snow-filled channels and flat, poorly-drained watershed areas draining to the channels differ significantly. It is shown that when overland flow from the watershed area to the stream channel is the major source of direct runoff and the unit runoff potential is reduced in an amount to account for other losses (primarily depressional storage), the volume of runoff can be related directly to the unit runoff potential and the "gross" drainage area.

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INTRODUCTION

A number of models have been developed for synthesizing streamflow from snowmelt. For example: the U.S. National Weather Service River Forecasting System, NWSRFS (U.S. Department of Commerce, 1972; Anderson, 1973; Peck, 1976); the Streamflow Simulation and Reservoir Regulation Model, SSARR (U.S. Army Corps of Engineers, 1972); the Hydrological Engineering Centre HEC-1 Flood Hydrograph Package (U.S. Army Corps of Engineers, 1973); the U.S. Department of Agriculture Hydrograph Laboratory Model USDAHL-74 (Holtan et al., 1975); the HBV of the Swedish Meteorological and Hydrological Institute (Bergström, 1978) and the UBC Watershed and Flow Model (Quick and Pipes, 1972). Of the systems which have been developed no model has been adopted for universal use. Similarly, in Canada the water management agencies of different Provinces responsible for streamflow forecasting are not unanimous in their choice of a model that best satisfies their requirements and produces the most reliable and accurate results. For example, recently the Ministry of Natural Resources, Province of Ontario tendered a study of the suitability of eighteen snowmelt subroutines used in existing, established models for interfacing with their streamflow forecasting system (Maclaren Plansearch, 1984). Ongoing tests are being conducted by the World Meteorological Organization and other bodies on the performance of a number of the "better-known" models in different physiographic and climatic regions of the world.

Each model differs from another, either as it calculates hydrological components or simulates the processes of snowcover accumulation and ablation, evaporation, infiltration, changes in soil moisture storages and flood routing. A main factor contributing to the differences is that most models were developed under a specific set of physiographical conditions, e.g. climate, topography, vegetation and soil types. Consequently, a model developed in a mountainous area will not usually give reliable streamflow simulations for a prairie watershed; nor should it be expected. Even when a model is applied to an area with similar characteristics to those where it was developed, extensive calibration and testing of its performance is usually necessary.

The simulation of snowmelt runoff has been most successful where there is abundant, uniformly-distributed snow and pronounced topographical relief. The opposite situation exists on the Prairies. The snowcover is shallow and large estimation errors in snowcover depth and water equivalent frequently occur, there is little relief and evaporation, infiltration and depressional storage are often comparable in magnitude with the total water content of the snowcover. Definition of a basin's snow resource is complicated by the interaction of wind with both topography and snowcover. This biases measurements and obscures the runoff potential. Continued strong winds erode friable snowcovers and transport both falling and erodible snow downwind. A comparison between measurements of snowcover at climatological stations and adjacent windswept fields showed that only about two-thirds of the snowfall recorded at the stations was retained on the fields (McKay, 1963). Much of the "lost" snow sublimates during wind transport or accumulates in shelterbelts, ditches and gullies. For that reason snow surveys taken in exposed sites can be very poor predictors of the runoff potential.
The performance of a model in simulating streamflow from natural catchments is directly related to the accuracy with which infiltration is evaluated. In most operational systems infiltration is estimated by empirical equations such as those reported by Horton (1940) and Holtan (1961), soil moisture accounting routines, or from relationships that index antecedent groundwater storage conditions and the soil moisture storage potential to the base flow recession characteristics of the streamflow hydrograph. Two main problems arise in applying these procedures to watersheds in northern and west-central Canada namely; (1) no attempt is made to distinguish differences in the infiltration process to unfrozen and frozen soils and (2) many streams are ephemeral, i.e. flow only occurs following a rainfall or snowmelt event and therefore the recession properties of the hydrograph do not properly index the soil moisture storage of a basin at the time of runoff.

Runoff from a Prairie watershed is not generated uniformly from the area enclosed by the topographical divide of the basin. Prairie lands are relatively-flat and their natural drainage systems are often poorly developed and unconnected. At times there may be no contribution of runoff from large areas of the watershed because of the lack of snowcover and the large amounts of depressional storage. Yet even under these conditions significant runoff can occur, the source being snow in the less visible channels and depressions that feed the main drainageway. A 1966 survey of in-channel snow near Regina showed 12,322 m$^3$ of water/1000 m of channel when the snow had virtually disappeared from adjacent fields and before significant streamflow. In 1984, surveys near Melfort, Saskatchewan showed an average snowcover of 540 mm in ditches adjacent to fields having an average depth of 70 mm. Snow in the ditches was denser than in the fields; an average density of 340 kg/m$^3$ compared with 190 kg/m$^3$ and contained more ice layers. Rough calculations suggested the water equivalent of the snow in the ditches of about 16,430 m$^3$/1000 m. In-channel snow, although an important source of water in low snow years, is an impediment to flow and the nature of its water storage, transmission and melt characteristics can be critical to hydrograph synthesis, particularly on small and medium-sized watersheds. As storage is filled overflow enters the channels. The result is that a Prairie basin has a variable "contributing" area whose magnitude varies with such factors as the amount of snowfall and antecedent soil moisture and surface storage conditions. In this regard hydrologists have made use of the concept of "Effective" and "Gross" drainage areas. The "Effective" area is that portion of a basin which might be expected to entirely contribute runoff to the main stream during a flood with a return of two years; the "Gross" area is the plane area enclosed by the drainage divide which might be expected to contribute runoff to the outlet under extremely wet conditions (Godwin and Martin, 1975). The "Effective" area includes the major channels and land immediately adjacent to defined drainageways. It is the snowcover in these areas that needs to be surveyed in low snow years; this component of the average basin snowcover becomes less important in snowier winters because of the larger area contributing to runoff. Further research is needed into the interaction of snowfall, snowcover and topographical aspects, and contributing area.

This following material focusses on the effects of infiltration to
frozen soils and "contributing" area as they affect the synthesis and determination of streamflow rates and volumes from snowmelt in a Prairie environment.

DEVELOPMENT OF INFILTRATION ALGORITHM

Previous discussions indicate that a major limitation of existing models in synthesizing streamflow from snowmelt on the Prairies is their inability to simulate the process of infiltration to frozen soils. Based on approximately fifteen years of study of the snow hydrology of the Prairie region, the results of infiltration studies in similar climatic regions of the USSR reported in the literature (Motovilov, 1978, 1979; Popov, 1973) and the findings of a comprehensive study of infiltration to frozen soils in the Dark Brown and Brown soil zones of Saskatchewan (Granger et al., 1984), it is suggested that frozen soils may be grouped into three broad categories with regard to their infiltration potential, namely; Restricted, Limited and Unlimited (see Fig. 1).

Restricted - infiltration is impeded by an impermeable layer, such as an ice lense on the soil surface or at a shallow depth in the soil. For practical purposes the amount of meltwater infiltration can be assumed to be negligible and most of the snow water goes to evaporation or direct runoff.

Limited - infiltration is governed primarily by the snowcover water equivalent and the frozen water (ice) content of the soil layer, 0-300 mm.

Unlimited - a soil containing a high percentage of large, air-filled, non-capillary pores or macropores at the time of melt and most or all the snow water infiltrates. Examples of these soils are dry, heavily-cracked clays and coarse, dry sands.

In the classification it is evident that when evaporation and surface storage losses are neglected the runoff coefficients to be assigned to soils of Restricted and Unlimited infiltration potential in a practical modelling scheme would be 1.0 and 0 respectively. Thus, the problem remaining is one of defining the relationship between infiltration, snowcover water equivalent and a frozen soil moisture content for the Limited case. This can be done using the results reported by Granger et al. (1984). They found in medium to fine-textured, uncracked frozen Prairie soils in which the entry of meltwater was not impeded by an impermeable layer that: (a) the average depth water penetrated a soil during snowcover ablation was 260 mm (standard deviation = 100 mm) and (b) the amount of snowmelt infiltration was inversely related to the average moisture (ice) content of the soil layer, 0-300 mm, at the time of melt. Based on these findings, they derived a set of equations defining the interrelationships between snowmelt infiltration (INF), snowcover water equivalent (SWE) and the premelt moisture content ($P_0$). For practical purposes, these results can be approximated by the equation:
(a) Restricted: Infiltration is low, high runoff potential.

(b) Limited: Infiltration is governed primarily by ice content of the soil layer 0–300 mm at the time of melt.

(c) Unlimited: Soil has the capacity to infiltrate all or most of the snowcover water equivalent.

Figure 1. Conceptual model for classifying the infiltration potential of frozen Prairie soils: (a) Restricted, (b) Limited and (c) Unlimited (after Gray et al., 1984).
\[ \text{INF} = 5(1-\theta_p)\text{SWE}^{0.59} \]

in which \( \text{INF} \) and \( \text{SWE} \) are in mm and \( \theta_p \) is the degree of pore saturation \( \text{cm}^3/\text{cm}^3 \).

Gray et al. (1985) have applied the infiltration model (Fig. 1) to calculate the runoff volume (areally-weighted snowcover water equivalent - infiltration = \( Q_{\text{cal}} \)) from the Creighton Tributary, a small watershed (11.4 km\(^2\)) in southwestern Saskatchewan. They compared the \( Q_{\text{cal}} \)-values with measured streamflow from snowmelt, \( Q_{\text{meas}} \), for 1974 and 1975; two years of different snowcover and soil moisture conditions and found a ratio \( Q_{\text{cal}}:Q_{\text{meas}} \) of 1.02 in 1974 and 1.16 in 1975. The results were considered sufficiently encouraging to incorporate the model into an operational system.

**Sequencing Infiltration Quantities - The Limited Case**

In order to apply the model in operational practice the variation in infiltration rate with time over the melt period must be assumed. The infiltration pattern depends on many factors: the rates of snowmelt and snowcover runoff; the depth, temperature regime and water transmission characteristics of the snowcover; the content and distribution of ice in the frozen soil, the soil temperature regime and others. Figure 2 (Granger et al., 1984) illustrates the effects of premelt soil moisture, snowcover depth and melt conditions on snowmelt infiltration.

![Figure 2. Snowcover infiltration curves. Curve 1: dry soil (-14\% moisture by volume), deep snowcover and variable rates of snowmelt; Curve 2: (-18\% moisture by volume), rapid melt of snowcover, and water ponds on the surface; and Curve 3: (-35\% moisture by volume), rapid melt and ice layer forms at the surface early in the melt period (after Granger et al., 1984).](image)
Curve 1 - infiltration to a relatively dry soil (\(<14\%\) moisture content by volume) resulting from the slow melt of a relatively-deep, Prairie snowcover (\(<500\) mm). Infiltration is delayed by the movement and storage of meltwater in the snowcover. After the snow ripens, water is released almost continuously throughout the melt period. Infiltration occurs at variable rates with a trend for the rate to increase with time because of an increase in the rate of release of meltwater and thawing of the soil.

Curve 2 - infiltration to a relatively dry soil (\(<18\%\) moisture content by volume) caused by the rapid melt of a ripe snowcover and water ponds on the soil surface providing a reasonably constant supply. The maximum infiltration rate occurs early in melt and the soil moisture storage, which is limited because of refreezing of meltwater in the soil and reduced downward percolation, is satisfied after approximately nine days of melt.

Curve 3 - infiltration to a relatively wet soil (\(<35\%\) moisture content by volume) resulting from the rapid melt of a shallow snowcover; an ice layer formed at the soil surface and prevented infiltration until it thawed on the fifth day of melt. The amount of infiltration is low.

In view of the strong dependency of infiltration on the melt process it would seem logical to allow the output of meltwater from the snowcover generated by the ablation subroutine of a model to be a dominant factor in sequencing the infiltration amount. Three approaches to the problem have been reported by Gray et al. (1985) namely: (1) all meltwater released by the snowcover was assumed to infiltrate until the infiltration potential of the frozen soil, as calculated by [1], was satisfied; (2) the infiltration rate throughout the period of melt was assumed constant - an approach similar to the infiltration-index method used in rainfall-runoff studies and (3) the amount of infiltration in a given period of melt (usually 6h or one day) was a constant, fixed percentage of the amount of meltwater produced in the time interval. The authors showed the effects of these different procedures on simulated hydrographs of snowmelt on a small watershed in western Saskatchewan. It was found that the simulations produced by (3) above were in reasonable agreement with observed hydrographs; the main problem with the method occurred with a delayed melt pattern. Runoff was generated in advance of active streamflow because infiltration was taken as a percentage of the amount of meltwater released by the snowcover and independent of the rate of release. To alleviate this problem the current study employs a modified procedure that makes use of methods (2) and (3).

A snowmelt infiltration index—the average rate of infiltration assuming a continuous, uninterrupted melt was used to establish the occurrence of direct runoff. Runoff does not begin until the rate of release of meltwater from the snowcover is greater than the snowmelt infiltration index. Gray et al. (1985) suggest an initial value for the "daily" snowmelt infiltration index equal to the snowmelt infiltration potential ([1]) divided by 6d. When the meltwater flux exceeds the index, direct runoff is calculated by multiplying the amount of meltwater produced, by the ratio of the snowmelt infiltration potential to the snowcover water equivalent at that time.
An algorithm of the infiltration model was interfaced with the U.S. National Weather Service River Forecasting System Sacramento Model (NWSRFS) and the system was used to synthesize streamflow from snowmelt on the Wascana Creek watershed at Sedley, Saskatchewan. The basin is located approximately 50 km southeast of Regina in the Dark Brown soil zone. Approximately 85% of the basin is under cultivation of cereal grains by dryland farming; the remaining area is in pasture, woody vegetation, roads, farmyards and townsites.

The topography of the area is flat to gently rolling and because of the poor relief and drainage development the portion of the "gross" drainage area that contributes to the annual streamflow from snowmelt can vary widely. Figure 3 is a schematic of the "effective" and "gross" drainage areas of the watershed; the size of each area was 236.7 km\(^2\) and 1634.3 km\(^2\) respectively (PFRA, 1985). Note, the "effective" area includes the main

![Schematic diagram of Effective and Gross drainage areas for Wascana Creek at Sedley, Saskatchewan.](image)
channels and parcels of land falling adjacent to defined drainageways. Streamflow at the Sedley gauging station is also affected by a small reservoir at Tyvan (see Fig. 3). Outflow from the spillway of the Tyvan dam lags the occurrence of the peak flow rate at Sedley and separation of discharge from this source from the observed hydrograph was accomplished by routine procedures. The effect of the dam on drainage area size is to reduce the "effective" area to \(125 \text{ km}^2\) and the "gross" area to \(350 \text{ km}^2\).

Data on soil moisture, snowcover depth and water equivalent and streamflow used in the analyses were obtained from Federal and Provincial Agencies namely: the Water Management Service and Atmospheric Environment Service, Environment Canada; the Hydrology Branch, Saskatchewan Water Corporation, and the Swift Current Research Station, Agriculture Canada (Davin Watershed). In 1981/82 these data were supplemented by snowcover and soil moisture measurements made on the watershed as part of a joint Canada-United States investigation of the application of airborne gamma techniques for estimating the areal snowcover water equivalent (Carroll et al., 1983; Goodison et al., 1984).

Temperature data for the snowmelt calculations and precipitation data used to update snow survey measurements after March 1 were from the Regina Airport.

RESULTS AND DISCUSSIONS

Comparisons of Observed and Simulated Hydrographs

Figure 4 shows the observed streamflow hydrographs for a low flow year, 1972 (Fig. 4a) and a high flow year, 1982 (Fig. 4b) plotted with three hydrographs generated with the NWSRFS: (1) LAND - the system was operated with its "LAND" subroutine and a contributing area of \(125 \text{ km}^2\), the "effective" drainage area; (2) Revised - the NWSRFS with the LAND subroutine replaced by the infiltration algorithm and a drainage area of \(125 \text{ km}^2\) and Revised, 144 (Fig. 4a) or 296 (Fig. 4b) - the NWSRFS with the infiltration algorithm and a drainage area (144 \(\text{ km}^2\) or 296 \(\text{ km}^2\)) which produced the observed volume of runoff. The respective return - period volumes and peaks correspond to approximately 3y and <2y events in 1972 and the 10y and 50y values in 1982. All simulated hydrographs shown in Fig. 4 used the same model parameters; only the size of the drainage area was changed. Also, for the comparisons the simulated hydrographs were positioned so that the start of runoff agreed with the time of beginning of observed flow. From the data in Figs. 4a and 4b it can be observed:

(1) The NWSRFS with its LAND subroutine grossly underestimates the observed volume of runoff. For example, in 1982 the observed runoff, expressed as an equivalent depth of water on a drainage area of \(125 \text{ km}^2\), was 153 mm compared with a depth of 2 mm simulated by the LAND subroutine.
Figure 4. Observed and simulated hydrographs for Wascana Creek at Sedley, Saskatchewan: (a) low flow year, 1972 and (b) high flow year, 1982.

LAND - the NWSRFS was operated with its "LAND" subroutine and a contributing area of 125 km², the "effective" drainage area; REVISED 125 - the NWSRFS with the LAND subroutine replaced by the infiltration algorithm and a drainage area of 125 km²; and REVISED 144 (Fig. 4a) and REVISED 296 (Fig. 4b) - the NWSRFS with the infiltration algorithm and a drainage area (144 km² of 296 km²) which produced the observed volume of runoff.
(2) The infiltration model significantly improves the performance of the NWSRFS in simulating streamflow from snowmelt over that obtained with the LAND subroutine. A measure of the improvement is obtained by comparing the observed and simulated discharge rates using the non-dimensional parameter $R^2$ defined by the equation (Nash and Sutcliffe, 1970):

$$ R^2 = 1 - \frac{1}{n} \sum_{i=1}^{n} (q_{oi} - q_{si})^2 / \frac{1}{n} \sum_{i=1}^{n} (q_{oi} - \bar{q}_o)^2 $$

where $n$ = number of values at evenly-spaced time intervals, $q_{oi}$ = observed discharge rate, $q_{si}$ = simulated discharge rate, and $\bar{q}_o$ = mean of the observed values.

$R^2$, termed efficiency, can be likened to the coefficient of determination used in statistics. The closer the positive value is to unity, the closer the agreement between observed and simulated hydrographs. Assuming a drainage area of 125 km$^2$ the $R^2$-values obtained with hydrographs generated with the LAND routine in 1972 and 1982 were -0.32 and -0.15, respectively, compared with $R^2$-values of 0.70 and 0.48 for hydrographs simulated with the infiltration algorithm.

Gray et al. (1984) have shown that the results above and in item (1) can be expected unless the LAND subroutine is adjusted to account for the differences between the infiltration characteristics of unfrozen and frozen soils. These adjustments involve limiting the water storage capacity of the Upper zone of the soil moisture accounting system and restricting percolation from the Upper zone to the Lower zone. Both of these changes can be physically rationalized and Gray et al. show they significantly improve the performance of the NWSRFS in simulating observed data.

The agreement between the shapes and time elements of the simulated hydrographs using the infiltration algorithm and observed hydrographs vary in the two years. For 1972 the agreement would be classed as fair to poor; in 1982 it would be considered satisfactory. Because the NWSRFS has a high degree of flexibility the "fit" could be easily improved by changing different parameters in the forecasting system. However, at this time such changes would lack a physical base. Previous work with the NWSRFS (Division of Hydrology, 1977) has shown that a major factor contributing to poor agreement between simulated and observed hydrographs when the system is applied for synthesizing streamflow from snowmelt on prairie watersheds is an incorrect simulation of the ablation process. For example, the manner in which the subroutine accounts for changes in the internal energy of a shallow snowcover; differences in the magnitude of the melt factor, liquid water holding content and water transmission
properties of the snowcover; the use of a temperature index for calculating melt and other factors. Some of the effects on the simulations were evident where it was found the time of simulated runoff consistently lagged the beginning of observed flow. An analysis of ten years of records showed delay times in the range of 2-5 days. In general, the larger values were associated with years of low snowcover and it is suspected that the trend may reflect a problem in accurately simulating the meltwater release and water transmission properties of snow-filled channels.

(3) In years of low flow the "effective" area (125 km²) of the watershed is the better estimator of the "apparent" area of the watershed contributing to the peak discharge; in years of high flow a contributing area equal to or less than the "gross" area (350 km²) gives the better simulation. A dependency in the degree of association between hydrographs on basin size is expected because the volume of snow water used as input to the system is directly related to area.

Interaction Between Runoff Potential, Runoff Volume and "Apparent" Contributing Area

Table 1 lists some general statistics on snowcover, soil moisture, infiltration, runoff and contributing area for Wascana Creek near Sedley, Saskatchewan for seven years in the period 1972-1982 inclusive. Data for 1973, 1977 and 1981 were not included because the snowcover in those years was exceptionally low; in amounts which would put in question the accuracy of the measurements of snowcover water equivalent and streamflow. Conversely, data for 1975, a year of high flow, were not listed because it was impossible to separate the contribution to streamflow at Sedley by outflow from the Tyvan reservoir.

In view of the narrow range in values of the premelt soil moisture content given in Table 1 (θp: 0.41-0.53) one is tempted to assume that the antecedent "moisture" conditions on the watershed, in terms of soil moisture and depressional storages, did not differ significantly between years. Under these conditions a strong association between the volume of observed streamflow and the unit runoff potential, the sum of the snowcover water equivalent (SWE) plus precipitation occurring during the melt period (PPT) less the volume of infiltration (INF) i.e., (SWE + PPT - INF) would be expected.

In 1972 and 1974, the average values of θp were 0.47 and 0.44 mm³/mm³, respectively (see Table 1). The small difference in θp-values occurred regardless of the fact that the summer and fall precipitation regimes in the preceding two years (1971 and 1973) differed appreciably. Rainfall on the basin from May to October totalled approximately 122 mm in 1971 and 400 mm in 1973; compared with a long-term seasonal average of 240 mm. In the two years there were also significant differences in the distribution of the rainfall and other regimes affecting the fall soil moisture regime. In 1971, approximately 30 mm of rain occurred five days prior to freeze-up and
Table 1. Annual statistics on soil moisture, snowcover infiltration runoff and contributing area for seven years of record on Wascana Creek near Sedley, Saskatchewan, 1972-1982.

<table>
<thead>
<tr>
<th>Year</th>
<th>Premelt Moisture Content (θ₀)</th>
<th>Snow Water Equivalent (SWE)</th>
<th>Ppt During Melt (PPT)</th>
<th>Infiltration Potential (INF)</th>
<th>Unit Runoff Potential (SWE+PPT-INF)</th>
<th>Observed Streamflow Volume 10⁻⁶ m³</th>
<th>&quot;Apparent&quot; Contributing Area km²</th>
</tr>
</thead>
<tbody>
<tr>
<td>1972</td>
<td>0.47</td>
<td>45</td>
<td>4.4</td>
<td>24.5</td>
<td>25.1</td>
<td>3.63</td>
<td>144</td>
</tr>
<tr>
<td>1974</td>
<td>0.44</td>
<td>137</td>
<td>0.0</td>
<td>49.5</td>
<td>90.0</td>
<td>24.00</td>
<td>267</td>
</tr>
<tr>
<td>1976</td>
<td>0.48</td>
<td>100</td>
<td>1.6</td>
<td>38.5</td>
<td>63</td>
<td>21.50</td>
<td>341</td>
</tr>
<tr>
<td>1978</td>
<td>0.41</td>
<td>42</td>
<td>0.4</td>
<td>26.4</td>
<td>16.8</td>
<td>3.00</td>
<td>188</td>
</tr>
<tr>
<td></td>
<td>0.41</td>
<td>60°C</td>
<td>0.4</td>
<td>32.4</td>
<td>28.3</td>
<td>3.00</td>
<td>106</td>
</tr>
<tr>
<td>1979</td>
<td>0.47</td>
<td>83</td>
<td>0.0</td>
<td>36.0</td>
<td>47</td>
<td>12.75</td>
<td>271</td>
</tr>
<tr>
<td>1980</td>
<td>0.49</td>
<td>51</td>
<td>0.0</td>
<td>25.4</td>
<td>26</td>
<td>4.25</td>
<td>163</td>
</tr>
<tr>
<td>1982</td>
<td>0.53</td>
<td>99</td>
<td>0.0</td>
<td>34.4</td>
<td>64.6</td>
<td>19.13</td>
<td>296</td>
</tr>
</tbody>
</table>

a) θ₀ = average premelt soil moisture (ice) content of the soil layer, 0-300 mm.

b) "Apparent" Contributing Area = drainage area of the watershed used in the simulations to produce the observed volume of flow.

c) SWE estimate biased to snow survey measurements in channels and depressions.
increased the moisture content of the 0-300 mm soil layer to a level higher than would be expected following a very dry summer. Conversely, the fall of 1973 extended through October and resulted in relatively-dry soil moisture conditions at freeze-up. These discussions suffice to point out the necessity of reviewing precipitation records close to the time of freeze-up in establishing the premelt moisture content. Gray et al. (1985) have shown over winters free of meltwater infiltration from midwinter melt or rainfall events that the premelt moisture content of the 0-300 mm layer is linearly related to the fall moisture content. The higher the moisture content in the fall the higher the moisture content in the spring and vice versa.

It is considered pertinent to point out to practicing modellers that the assumption of a high correlation between the moisture content of the 0-300 mm soil layer and available depressional surface storage is most likely in very wet years. No measureable surface runoff was recorded in 1971 from the Davin watershed, a small 12 km² watershed located about 27 km north of the Sedley gauging station, whereas in 1973 a 125-mm rain on June 30th produced significant surface runoff and high discharge rates; measureable surface runoff was also recorded in September and October. Accordingly, although the soil moisture conditions in the two years were similar the amount of depressional storage available to collect and retain direct snowmelt runoff would likely be larger in 1972 than in 1974. Note, the relative effects of depressional storage on streamflow would be smaller in 1972 because most of the streamflow would likely originate from the snow-filled channels.

Figure 5 is a plot of the observed volume of streamflow with the unit runoff potential for the seven years of data given in Table 1. The numbers opposite the plotted points represent the "apparent" contributing area, the area of the basin that produced the observed runoff volume with the unit runoff potential. The data in the figure show a trend for the volume of runoff to increase linearly with the unit runoff potential which is relatively-independent of the "apparent" contributing area; a trend that was unexpected considering the storage elements between years may have differed appreciably.

In the figure three line segments have been plotted and numbered. The primary division of the data (Curves 1 and 3) is rationalized on the basis that the runoff response characteristics of poorly-defined drainage areas of the watershed adjacent to the channels differ significantly from those of the snow-filled channels. That is:

Curve 1 - runoff from the "gross" area of the basin assuming the surface storage elements and the snowmelt infiltration potential of the frozen soils are satisfied; infiltration occurs at a very low and reasonably-constant rate; and evaporation losses are negligible. The line has a slope equal to the "gross" area of the basin or 350 km², and represents the upper "envelope" curve of runoff for the basin.

Curve 3 - runoff from snow-filled channels. The runoff relationship is approximated by a curve having a concave shape showing an increase
in runoff per unit runoff potential with increasing potential. Although the shape can not be verified it is rationalized on the basis that as the amount of snow water collected in the channels increases the rate of runoff released per unit of snow water equivalent also increases, until such time when direct runoff contributions from areas of the watershed adjacent to the channels become the major source of supply.

Curve 2 can be likened to a transition zone in which the "apparent" area contributing to streamflow enlarges in size according to the magnitude of the runoff potential and reaches a maximum value equal to the "gross" area. In this segment the association between observed streamflow and runoff potential is assumed linear because the relationship defined the trend of the data within acceptable, practical limits and there was no obvious physical reason for describing the relation with another geometric form. Also, no attempt was made to determine the best-fit regression for the data because of the small number of data points and the values of the unit runoff potential were calculated from measurements which were routinely-collected as part of an operational field survey program. That is, the
measurement program was not specifically designed to warrant a rigorous, statistical approach. An important property of the line is that the slope is equal to 350 km$^2$, the "gross area".

Two other pertinent properties of Curve 2 (Fig. 5) are the intercepts with the "x" and "y" axes and the point of intersection or tangency of the line with Curve 3. The "x" intercept represents the unit runoff potential above which the volume of runoff can be assumed a linear function of the "gross" area. Thus, it can likened to a displacement value, similar to a roughness height used to define wind velocity and other profiles. The "y" intercept (vertical displacement from Curve 1) represents the average volume of storage (primarily depressional) that must be satisfied before runoff from the watershed can be directly related to the "gross" area. For Wascana Creek this value was approximately 3.9 $\times$ 10$^6$ m$^3$. Note, like Curve 1 which represents the "upper" envelope curve of maximum runoff for a given potential, a line positioned some distance vertically-below and paralleling Curve 2 would describe the runoff characteristics under maximum surface and depressional storage characteristics, i.e. the case of very dry antecedent moisture conditions.

It is assumed the point of tangency of Curves 2 and 3 represent the "apparent" contributing area when the runoff release and response characteristics change from "channel" to "watershed" patterns. The area was calculated in the range of 150-160 km$^2$, a size slightly larger than the "effective" area of 125 km$^2$.

The findings above suggest procedures for estimating: (1) the runoff volume from snowmelt from the runoff potential and the "gross" area of the basin; (2) an initial value of the "apparent" contributing area for use in model simulations and (3) the size of a watershed when its runoff response changes from "channel" to "watershed" characteristics. At the present time these findings lack a strong physical base, therefore they must be accepted as empirical. However, the simplicity of the procedures and relationships and their practical application warrant additional study and investigation.

SUMMARY

The paper discusses the effects of infiltration to frozen soils and the area of a watershed contributing to flow on the simulation of streamflow and the estimation of the volume of runoff from snowmelt on Prairie watersheds. An infiltration model for frozen soils is presented and methods of interfacing it with an operational streamflow forecasting system, the U.S. National Weather Service River Forecasting System (NWSRFS), described. It is shown that an algorithm of the model in the NWSRFS significantly improves the performance of the forecasting system in simulating streamflow from snowmelt from the Wascana Creek basin at Sedley located approximately 50 km southeast of Regina, Saskatchewan. Comparisons illustrating the improvement between simulated and observed hydrographs for both "low" and "high" runoff years are presented.
A method for estimating the volume of runoff from a prairie watershed based on the unit runoff potential, the sum of the snowcover water equivalent plus precipitation during the ablation process less the amount of snowmelt infiltration, and the "gross" area of the basin is demonstrated. Based on 10 years of field data for Wascana Creek it was found the association between variables is approximately linear, provided the runoff potential (or runoff volume) is reduced in an amount to account for other losses, e.g. depressional storage. The analyses are based on the assumption that the runoff response and release characteristics of snow-filled drainageways differ appreciably from those of poorly-defined watershed areas adjacent to channels.

In addition to providing a means for estimating the volume of runoff from snowmelt the results suggest procedures which could be used in differentiating runoff from channel and watershed areas and for determining an "initial" value of the size of the contributing area of a watershed to be used in hydrograph simulations. At the present time the findings lack a complete physical base and therefore must be treated as empirical. However, the simplicity of the procedures and the practical application of the results suggest additional research on the concepts is warranted.

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REFERENCES


