MELT OF SHALLOW PRAIRIE SNOWPACKS: BASIS FOR A PHYSICAL MODEL

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Prediction of the volume of water available for runoff and the rates of runoff due to snow melt necessitates (a) a determination of the depth and density profile of the pack, (b) a knowledge of the energetics of the melt process, and (c) a knowledge of the accompanying infiltration process. In each of these areas, shallow packs and deep mountainous packs have common characteristics; however, in many ways they differ. The greatest differences occur in the energetics of the melt process and in the infiltration process. For example, it has been found that the albedo values decay rapidly with time during the melt period of shallow packs as compared to those of deep packs. In shallow packs the distribution of the radiant solar energy throughout the pack depends a great deal upon the depth of the active layer. In many cases the active layer is actually the entire pack; consequently many of the "deep pack" snowmelt models are unsuitable for predicting shallow pack melt. Infiltration often accounts for the major portion of the melt water from shallow packs; however, its significance depends upon the condition of the soil at the time of freeze-up. In general the soil below shallow prairie packs is below 0°C when melt commences. If the soil was near saturation when freeze-up occurred the infiltration rate may initially be very small.

INTRODUCTION

Traditionally, snow hydrology studies have been associated with the deep packs of mountainous regions. These studies have generally been conducted for two purposes: (1) to determine the total volume of water available for runoff, and (2) to estimate the peak runoff rate. Knowledge of the total volume available is important from the standpoint of water supply to major urban areas for both domestic and industrial purposes. This source may also serve the irrigation requirements of adjacent areas. An estimate of the peak runoff rate is obviously necessary for flood forecasting. It is natural then that studies have been concentrated on these two main aspects, water supply and flood forecasting.

More recently it has been realized that the shallow packs of the Canadian Prairie and Arctic regions are also important and worthy of detailed studies. The snow resources of these areas have a great deal of impact on the people and environment of the area. On the prairies the snowpack is generally the main source of potable fresh water supply for local domestic use. In addition, the peak runoff from these shallow packs can cause local flooding of significant proportions although in general these are not of the spectacular nature of mountain runoff floods. From these standpoints, runoff from both mountain packs and prairie packs have similar effects on the population although the magnitude, in terms of numbers of people affected, is different. However, the snow resource of the Prairie region has additional direct impact on the population, the greatest factor probably being that of replenishment of soil moisture for crop production. From this standpoint, studies show that improved management techniques can result in increased soil moisture storage, the resulting increased crop production would be of considerable direct monetary benefit to the area. Other examples of problems concerned with snow conditions which persist for several months of the year relate to transporation and the effects of soil moisture and soil temperature regimes on construction works.

In general, the investigations of mountainous snowpacks have been directed to the development of models for predicting or forecasting streamflow or discharge. Their use in the prairie environment is highly questionable because of basic differences in the climatic, vegetal and topographic features which cause differences in the snow hydrology regime of this region. Further, as suggested previously, although runoff rates and volumes are important, a prairie model must also be capable of predicting other physical parameters such as soil moisture and soil temperature. Consequently the development of a prairie snowmelt model should be based on a good understanding of the physical processes involved. For this purpose, the energy budget approach — which simply involves accounting for the thermal energy involved — is an appropriate framework within which to develop a model. Also it is expected that such a model may be transposed for use to study snowmelt problems in other parts of the continent having similar topographic and climatic conditions (i.e., Arctic).

DEPTH-DENSITY MEASUREMENT OF SHALLOW PACKS

Depth and density measurements of shallow prairie snowpacks are complicated by the effects of wind. The average wind velocity in the prairie region during winter months is 4 - 6 m/sec⁻¹ (15 - 25 km/h⁻¹) with maximums of approximately 20 m/sec⁻¹ (75 km/h⁻¹). When snowfall occurs under such conditions it is extremely difficult to obtain meaningful measurements with conventional precipitation gauges. Black (1954), Gray et al. (1970a) and Peck (1972) all point out that under windy conditions, standard precipitation gauges undercatch when compared to ground survey measurements. Gray et al. found that Fischer and Porter precipitation gauges on the average only registered 43% of that measured on the ground.

If ground surveys are used to establish the water equivalent of snowpacks, difficulty is also encountered due to the effects of wind. On the prairies, although the general snowpack depth is fairly uniform, local variations in topography and vegetative cover may cause major departures from this average. Each field, therefore, has its own peculiar catch and retention characteristics. This fact, of course, has long been recognized by agriculturalists in that strip-cropping is used as much as a moisture conservation practice as a wind erosion control measure. Studies have shown that stubble fields retain as soil moisture, on the average, an amount of water equivalent to approximately 37% of that of the average overwinter pack whereas fallow lands retain only about 9%. Part of this difference can, of course, be attributed to
the fact that the stubble will retain snow blown from adjacent fallow strips which under certain conditions may be completely denuded.

In general, because of the expanse of areas of flat or gently rolling topography, the sparsity of tall, dense vegetative growth and the continual strong, surface winds, severe drifting and redistribution of the snowpack may occur over the winter months.

Gray et al. (1970b) reported that depending on wind conditions and the time of measurement after snowfall begins, the densities of freshly fallen snow measured at the ground surface vary in the range from 0.04 to 0.23. It was also found that because of wind abrasion, packing and other factors, freshly fallen snow quickly reached a density in the range of the average density of the over-winter pack, 0.25 - 0.30. Only in cases where sufficient vegetation was present to shelter and support the snow was the density found to be significantly less than the values measured in the open fields. Slaughter et al. (1973) reported drifted snow of density greater than 0.50 in the tundra region of Alaska. Billelo (1966) suggested that seasonal snowcover density could be related to air temperature and wind velocity, with the density decreasing with increasing average seasonal air temperature and increasing with average wind speed.

The use of twin probe gamma radiation snow gauges on shallow packs (Gray et al. 1970a) does allow the pack to be profiled for changes in density. However, unlike the case of deep mountain packs, since prairie snowpacks are extremely heterogeneous in their distribution it is nearly impossible to establish “representative” point sampling sites. Consequently, portable equipment must be used and many sites may have to be sampled to obtain a representative density profile.

Remote scanning of natural gamma radiation from the soil through the snowpack may prove to be useful in overcoming the problem of point sampling. Bissell and Peck (1973) monitored the natural gamma radiation from the soil with a detector placed 2 m above the ground surface. They found that the water equivalent could be determined with a standard error of 11 mm when the pack had a water equivalent of 50 - 400 mm.

Dmitriev et al. (1972) used natural gamma radiation methods to determine the basin water equivalent from aircraft. For packs with a water equivalent of 10 - 300 mm, the standard deviation of the measurements was less than 10 mm. They point out that as the radiation is originating from the upper 300 - 400 mm of soil, this process “sees” water which is stored as ice lenses on the ground surface, whereas conventional sampling methods generally do not measure this quantity.

Grasty et al. (1973) report the use of gamma ray spectrometry surveys using total radioactivity and potassium activity from natural sources to determine snow water equivalent. With presnow flights to obtain background counts, the major sources of error appear to be duplication of flight track, changes in soil moisture and differences in atmospheric pressure. For shallow packs (maximum water equivalent of 140 mm) the statistical errors due to low count rate were not significant. After corrections were made for soil moisture changes, the standard error for the potassium scan was 12 mm.

Linlor (1972) has shown that airborne electromagnetic wave methods can be used to determine the density profile of a snowpack. The work would be a step beyond those methods that only give a measure of the total water equivalent.

No doubt in the future, snow scanning by satellite methods will increase in importance, but as yet the most valuable use of satellite imagery is to define areas covered by snow rather than provide measurements of density or water equivalent. However McGinnis (1972) has shown that near-infrared data when used in conjunction with reflected visible radiation appear to be useful in detecting melting snow and ice. Under melting conditions the near-infrared radiation is strongly absorbed, while the visible radiation is reflected. Linlor (1972) speculated that satellites might be used for the electromagnetic wave methods.

For studies of shallow packs in the Prairie region, the use of satellite information in hydrologic studies, particularly in the study of runoff events, may be limited because of the short time period over which melt occurs. Presently, the time interval between consecutive passes over an area is several days and thus the system may miss the most significant time of the melt season.

THE ENERGY BUDGET

As stated previously, it is considered that the most feasible method of developing a snowmelt model for the Prairie region is using the energy budget approach, as it is based on the actual physical processes. Carlson et al. (1972) have pointed out the necessity of a model having as much physical reality as possible to permit transposition of the results to other areas and to make full use of sparse input data. In addition, they point out that the model should have as few empirically-derived parameters as possible to allow full latitude for improvement of the model.

The energy budget approach can be described as equating the time rate of change of the energy of the pack to the energy fluxes into the pack; that is

\[
du/dt = QSS + QSL + QSH + QSM - QGS - QGC - QGM \tag{1}
\]

where

\[
U = \text{energy of the pack}, \\
t = \text{time}, \\
QSS = \text{net solar radiant flux into the pack at the snow-air interface}, \\
QSL = \text{net long wave radiant flux into the pack at the snow-air interface}, \\
QSH = \text{sensible heat flux into the pack at the snow-air interface}, \\
QSM = \text{net energy flux into the pack due to mass transfer at the snow-air interface}, \\
QGS = \text{net energy flux into the pack due to solar radiation through the pack at the ground beneath}, \\
QGC = \text{heat flux by conduction from the pack to the ground beneath}, \\
QGM = \text{net energy flux of snow due to mass transfer at the ground-snow interface}.
\]

From a thermodynamic standpoint, equation 1 is not complete, for it does not include terms for mechanical work at the snow-air interface, for work arising from the change in depth of the pack, for potential energy or for kinetic energy. However, it can be shown that each of these terms is negligible when compared to the accuracy with which the various terms of equation 1 can be measured.

On the prairies, the general pattern of snowpack disappearance is that snow first disappears from the fallow fields. This is followed by a gradual shrinkage of the remaining snowcover until only patches or drifts persist in gullies and sheltered areas. Melt from these deep residual drifts frequently is not evident as surface runoff until after the open fields are free of snow. Under these conditions the energy budget must be handled differently. When the ground is completely snow-covered the uniform underlying surface at the base of the atmosphere permits meteorological turbulence theory to be applied to the calculation of QSH and QSM. When the land is partly snow-free, the application of standard meteorological techniques for calculation of QSS, QSG and QSM is probably invalid, since the underlying surface is patchy and therefore nonuniform and heterogeneous with respect to the heat transfer processes.

Solar Radiation ($QSS$, $QGS$)

The incoming solar radiation at the earth's surface is generally considered to be a function of the solar altitude, the extent, distribution and form of cloudiness, the absolute humidity and the amount of ozone and dirt in the atmosphere. In addition, the amount of solar radiation received by a surface will
depend upon its slope and aspect. Gray and O'Neill (1973) have shown that this is important during the melt process of shallow prairie packs. They investigated the energy exchange of different slopes of a prairie watershed during 6 days of the 1972 snowmelt period. Their findings indicated that by simply adjusting the direct beam component of incoming shortwave radiation, the net radiation to a south-facing slope was approximately five times greater than the amount received by a similar north-facing slope. South-facing slope contributed essentially no runoff when a pack with 46 mm of water equivalent was on bedrock. It appeared that the entire pack was lost to sublimation and evaporation. A similar north-facing slope yielded 82% runoff. On vegetated areas (5-m jackpine) approximately 70% runoff occurred on both slopes, although the south-facing slope yielded greater peak flows.

The reflected solar radiation is related to the incoming radiation by the albedo factor. Studies conducted by O'Neill and Gray (1972a) showed that during the melt-free period the albedo of prairie packs ranged from 70 to 80% depending on snowfall conditions. During the melt period, the time decay of the albedo showed an accelerated rate of change with time, quite dissimilar to the shape of the relationship usually assumed for deep packs. It was also found that point measurements of albedo were in close agreement with spatially averaged measurements obtained by flying over the snowpack (provided snow remained with the field of view of the sensor). O'Neill and Gray (1972b) found that the albedo and the extinction of solar radiation in snow are coupled and largely controlled by the properties of a thin "active layer" at the snow surface.

Solar radiation to the ground \( Q_{GS} \) has generally been assumed to be negligible or included in the ground heat flux term (Anderson 1968; Boyd et al. 1962; U.S. Corps of Engineers 1956). However O'Neill and Gray (1972b) have shown that this is often not true for shallow pack conditions. Their conclusion was that the radiative heat flux through snow during the melt season may be of significant magnitude for snowpack depths up to 100 mm. In addition they found that the simple diffusion model (Giddings and LaChappelle 1961), which describes radiation penetration in snow, when extended to the multivavefractional wavelength situation appears to seriously underestimate the solar radiation penetration to an absorbing surface below the active layer.

Net Longwave Radiation \( Q_{LSL} \)

In the absence of significant forest cover, such as those conditions encountered on the prairies, the principal factors influencing the longwave exchange are: (1) the temperature of the snow surface and air layer close to the ground; (2) the absolute air humidity; (3) the amount and form of cloudiness; and (4) the wind velocity in the air layer close to the ground.

Winters on the prairies are characterized by lengthy nocturnal periods of cloudless skies. During these periods, because of the low absolute humidity of the air, the longwave radiation loss is large and the total net radiation exchange is negative. That is, the outgoing radiation loss during the evening exceeds the gain during the day. Under these conditions the temperature of the pack is lowered.

Even during the melt period, the nocturnal radiation losses may be sufficient to refreeze all or a portion of the thawed soil and to reduce the temperature of the surface crust below 0°C. Usually, however, this loss does not exceed 15% of the average daily heat input. As the net longwave radiation exchange is dependent on both air temperature and humidity, this exchange can be greatly altered by a change in magnitude of these variables. It is a generally recognized fact that on the prairies, appreciable melting of the pack will not occur until the mean daily air temperatures exceed 5°C. It should be pointed out that at the time of melt there is often a reversal in the time of occurrence of maximum cloud cover. That is, the days often clear whereas clouds form in the evenings (probably caused by evaporation during the day). The obvious effect of the increased cloudiness is to reduce the nocturnal radiation loss and subsequent cooling of the snowpack. Since the shallow prairie pack responds quickly to daily temperature changes, these conditions are conducive to high melt rates.

With respect to radiative components it is considered that the immediate needs concern study of the development of methodology, techniques and procedures which will enable: (1) extrapolation of point radiation measurements in space and adjusting these according to a "gross" topographic and landscape model, and (2) accurate partitioning and evaluation of the radiative terms as they contribute to snowmelt, particularly under conditions of patchy snowcover.

Energy Transfer by Mass Flux at Snow-Air Interface \( Q_{SM} \)

The energy transfer at the snow-air interface due to mass flux is

\[
Q_{SM} = M_{SL} h_{SL} + M_{SV} h_{SV} \quad (2)
\]

where

\[
M_{SL}, M_{SV} = \text{liquid and vapor mass fluxes}\]

respectively, at the snow-air interface,

\[
h_{SL}, h_{SV} = \text{liquid and vapor enthalpies}\]

respectively.

The liquid flux into the pack, \( M_{SL} \), can be rainfall or condensate. The vapor flux, \( M_{SV} \) (usually negative), includes all forms of evaporation, sublimation, and vapor transfer. The term \( M_{SV} h_{SV} \) is commonly referred to as the latent heat transfer. This term is best discussed in conjunction with the sensible heat transfer process.

"Sensible" and "Latent" Heat Transfer \( Q_{SL}, M_{SV} h_{SV} \)

The sensible and latent heat transfer fluxes are generally calculated from the respective temperature and vapor pressure gradients. Various aerodynamic formulae such as those of Sverdrup (1936), U.S. Army Corps of Engineers (1956), Thornthwaite and Holtzman (1939), Dyer (1965), and Bowen (1926) can be used for these calculations.

Diamond (1953) has pointed out the misconception that appreciable evaporation takes place when a warm dry air passes over a snow surface. Since evaporation can only take place when a vapor pressure gradient exists (except for molecular diffusion) and since the maximum temperature a snow surface can reach is 0°C, the relative humidity of the warm air will have to be appreciably less than 100% if evaporation is to take place. In addition, a heat supply must be available for evaporation or the temperature of the snow surface will be lowered causing a subsequent reduction in the vapor pressure at the surface, and possibly condensation.

A portion of the Southern Canadian Prairies receives warm dry winds of relatively high velocity several times during the winter. These winds, known as Chinooks, may vary in duration from a few hours to several days and are usually accompanied by abrupt temperature changes of as much as 30°C. These winds may cause appreciable melting of the pack due to turbulent transfer processes as sensible heat transfer; however, it is doubtful that they will cause direct evaporation and/or sublimation. The disappearance of the pack under such conditions may be attributed to melting accompanied by infiltration and evaporation from a free water surface.

Gray and O'Neill (1973) have shown that under complete snow cover conditions on the prairies net radiation accounted for 93% of the total energy supply, while sensible heat contributed
7%. However, on an isolated snow patch, net radiation contributed 56% and sensible heat transfer supplied 44%, thus supporting the argument that as the snowpack melts and becomes patchy, significant amounts of heat are advected from snow-free areas and are used to melt the snow on the adjacent snow covered areas.

Heat Flux at Ground-Snow Interface ($Q_{GC}$)

During most winters on the prairies, the flow of heat within the ground underlying the snowpack is toward the soil surface and thus there is a gradual lowering of soil temperature. In the absence of the occurrence of an early large snowfall, depths of frost penetration of 2 m are common. However, because of the low thermal conductivities of the soil (usually at low moisture content), the relatively small thermal gradients, and the presence of the soil-snow interface, it is questionable whether the net transfer is of sufficient magnitude and rate to cause melting of the pack. Most likely this heat partially offsets the net loss through longwave radiation and thereby resists lowering of the temperature of the pack.

Energy Transfer by Mass Flux at Ground-Snow Interface ($Q_{GM}$)

Similar to the energy transfer at the snow-air interface, the energy transfer at the ground-snow interface due to mass flux can be written as

$$Q_{GM} = M_{GL} h_{GL} + M_{GV} h_{GV} \quad \ldots \ldots \ldots (3)$$

where

- $M_{GL}$, $M_{GV}$ = liquid and vapor mass fluxes respectively at the ground-snow interface,
- $h_{GL}$, $h_{GV}$ = liquid and vapor enthalpies, respectively.

The liquid mass flux, $M_{GL}$, actually includes the mass flux into the soil (infiltration) and the mass flux leaving the pack as runoff.

Although the vapor transfer across the ground-snow interface during the winter months may be significant, its contribution to energy transfer during the melt period is probably negligible compared to the contribution of the liquid transfer.

Energy of the Pack ($U$)

The energy of the pack can be written as

$$U = \int_0^L (\rho_i U_i + \rho_l U_l + \rho_v U_v) \, dx \quad \ldots \ldots \ldots (4)$$

where

- $L$ = depth of the pack
- $\rho_i$, $\rho_l$, $\rho_v$ = mass of ice, liquid and vapor respectively, in a unit volume of snow
- $U_i$, $U_l$, $U_v$ = ice, liquid and vapor specific energies, respectively.

When continuous melt is in progress, a snowpack rapidly reaches an isothermal condition at a temperature of 0°C (U.S. Corps of Engineers 1957). For deep packs, the total free water content probably becomes reasonably constant; therefore, the time rate of change of the energy of the pack can be approximated by the product of the time rate of change of depth, the density of the ice phase and the energy of the ice phase. In general, this value will be small compared to the various fluxes.

Under prairie conditions, the melt process is generally not continuous because of the shallowness of the pack (small energy storage capacity) and the radiant cooling at night. Thus, the change in internal energy becomes an important factor and has a marked influence on the diurnal fluctuation in the runoff pattern. It is well known that under prairie snowmelt conditions peak melt and runoff rates occur on days when the pack has not had an opportunity to refreeze during the preceding night. As a general rule, flooding from these shallow packs will not occur unless the overnight temperatures are above ~4°C.

For prairie conditions, equation 4 can best be evaluated from snow temperature measurements, a measurement of the average density of the pack and a determination of the free water content, probably by the calorimeter method.

**MELT WATER FLOW THROUGH THE PACK**

The energy budget approach is a useful concept which can be used to equate the thermal energies involved in the melt process; however, additional equations are required to describe the liquid flow process that takes place within the pack.

Colbeck (1972, 1973) and Colbeck and Davidson (1972) have treated the snowpack as a porous medium for purposes of describing water flow in the pack. The snowpack is considered to be made up of two layers, an upper layer of unsaturated snow wherein the flow is essentially vertical, and a lower layer of saturation in which the flow is basically horizontal unless infiltration is significant. Colbeck's method requires a measure of the density and liquid water content as a function of depth at some time, to provide an initial condition for calculation of flow through the unsaturated layer. In addition, the water flux across the surface must be known. From this standpoint it is clear that the energy budget and the flow process must be coupled mathematically.

Colbeck (1973) points out that the wave speed for flow in the unsaturated layer is significantly less than that in the saturated layer. Consequently, for deep packs the unsaturated zone will largely determine the delay in runoff. However, for shallow packs the timing of the runoff may be controlled by the saturated layer at the base.

**INfiltrATION INTO FROZEN SOIL**

For deep mountainous snowpacks, infiltration is often considered to be unimportant; however, for shallow prairie packs a major portion of the melt water may infiltrate into the soil. The volume infiltrated obviously depends upon the melt rate and the soil type, but the major factor controlling infiltration is the soil moisture status at the time of freeze-up in the fall. Murray and Gillies (1971) found that under prairie conditions there was a linear decrease in the amount of infiltration with increasing soil moisture content. In addition, the soil moisture content influences the shape of the infiltration rate curves of a frozen soil. These may adopt several distinct forms:

1. An intake rate which is reasonably constant with time at a very low value, a condition which would prevail if frozen while at a high moisture content or if an impervious layer develops at the surface due to refreezing of the melt water at time of thaw.
2. An intake rate which decreases very rapidly with time from a reasonably high initial value to near zero, a condition which may prevail when a soil is frozen at a low moisture content but the soil temperature is below freezing. Melt water entering the soil is frozen in the pores and movement is inhibited.
3. An increase in infiltration rate with time, a condition which may exist when the soil is frozen at an intermediate moisture content. In this case, some of the melt water is able to penetrate the soil and as the soil warms and more pores melt, the infiltration rate increases.

Obviously, the infiltration process is coupled to the conduction of heat in the soil profile because of the interrelationship between the soil temperature and the mass flow properties of the soil. It is conceivable that to fully understand the significance of this coupling, it may be necessary to compare many solutions of the coupled heat and mass transfer equations for frozen soil. Harlan (1973) has developed a numerical technique for obtaining such solutions.

**SUMMARY**

The snowmelt event is a complex phenomenon involving many physical processes. It is apparent that a snowmelt
model for the prairies must be based on an understanding of the energy transfers involved, the mass flow process within the pack and the heat and mass flow in the soil beneath the pack. It is hoped that an increased knowledge in each of these areas may contribute to the development of a simplified model rather than to increase the complexity of existing models.


