ENERGY TRANSFER IN SNOW

Don M. Gray and D. H. Male
Division of Hydrology
College of Engineering
University of Saskatchewan
Saskatoon, Saskatchewan

Introduction

During the past several years the Division of Hydrology, University of Saskatchewan, has carried out an extensive field program in Snow Hydrology. One group of studies in this program has been directed to investigation of the energy regime of the snowpack during both "non melt" and "melt" events. Consideration has also been given to the spatial and temporal variations of some of the energy terms and the transposability of data collected at a point to areal conditions. The data base established includes measurements taken from manually-obtained samples of natural snow, a snow lysimeter installed in the field which is capable of automatically monitoring most of the energy terms at short time intervals as programmed by a HP 2114B digital computer and measurements taken from the air.

It is recognized that the presence of a snow layer on an ice surface influences the thermal regime and energy transfer processes of the ice-water matrix thereby affecting such factors as: the rate of growth of ice, its strength properties and its decay or disappearance. In this abstract, the energy terms important to the thermal regime of a snowpack are discussed.

The Energy Equation

Consider the energy equation for a snow pack in the following form:

\[ \frac{dU}{dt} = Q_N + Q_H + Q_E + Q_M + Q_G \]  \hspace{1cm} (1)

where \( \frac{dU}{dt} \) = the rate of internal energy change per unit area (watts/sq. meter),

- \( Q_N \) = the net radiation flux at the snow-air interface,
- \( Q_H \) = the flux of sensible heat at the snow-air interface,
- \( Q_E \) = the flux of latent heat at the snow-air interface (evaporation, condensation),
- \( Q_M \) = the flux associated with melt water leaving the bottom of the snow pack, and
- \( Q_G \) = the heat flux at the snow-ground or snow-ice interface.
In any experimental study of the energy exchange for a snow pack it is important that each of these terms be measured independently so that an estimate can be made of the accumulated error inherent in the measuring procedures. Figures 1 and 2 are typical plots of the terms in Equation (1) over a 24-hour period in the Spring of 1974 when melt first occurred. On this day the snow depth was approximately 30 cm. Figures 3 and 4 are similar curves for a day in which no melt water was produced.

Radiative Terms

The term \( Q_N \) of Equation (1) represents the net all-wave radiation flux to or from the snowpack. Its magnitude is affected by the short-wave radiation incident to the surface, the albedo of the surface and the net long wave radiation exchange. On the prairies it has been found that this flux is extremely important and dominates the thermal regime of the pack during non-melt periods and also the melt phenomenon during the period when snow cover is complete. During the latter part of the melt sequence when the snow cover is patchy sensible heat becomes a dominant factor. The important parameters affecting this term are discussed below.

Albedo

Albedo, \( A \), is the ratio of the reflected, \( Q_{RS} \), to incident, \( Q_{IS} \), short wave radiation. It depends on atmospheric and surface conditions and the properties of the snow. Manz (1974) reported the absorptivity of a snow pack is greatly affected by the presence of foreign matter content (material other than ice, water or air) primarily because of the effects of such material on the extinction coefficients and albedos. Similarly, his findings indicated that the albedo characteristics of a snow pack decrease rapidly with an increase in density and particle size.

Figure 5 shows the temporal decay curves of spatial albedo (within the wave length band 0.2 \( \mu \)m to 1.2 \( \mu \)m) measured during 1974 over a lake surface and open fields in Saskatchewan. The curves show three distinct characteristics:

1. High albedos during the non-melt periods; 70 to 75 percent.
2. A rapid decay of the albedos of both surfaces during the melt period. Melt on the land surfaces started on April 8th.
3. A close association between the albedos of the two surfaces. Major differences between the two curves occurs only during the melt period. The more rapid decrease in the lake albedo is likely caused by ponded surface runoff and the shallower depth of snow cover on the lake surface. Water began ponding on the lake surface on April 18th. By April 24th the land was mostly free of snow and water covered the ice on the lake surface. The result accentuates the importance of snow cover in governing the albedo of a surface.
Other theoretical and experimental studies conducted have indicated that:

1. The underlying boundary surface only exerts measurable influence on the surface albedo when the depth of snow is very shallow (~2 - 4 cm).

2. The water content of snow exerts little influence on its optical qualities providing if it is not turbid and does not cause extensive modifications of the snow structure.

Flux

The flux of monochromatic diffuse radiation at any depth within a snowpack is characterized by the extinction coefficient, absorptivity and the speed of light. All these parameters are mathematically related. The extinction coefficient and albedo are highly wavelength dependent. The extinction coefficient exhibits a general increase at wavelengths greater than the near-infrared (> 0.7 μm) whereas the albedo decreases in this wavelength interval. O'Neill and Gray (1973) and Manz (1974) have shown that the broad-band radiation flux profile can be defined by the simple exponential function

\[ F = F_0 \exp(-bz) \]  \( (2) \)

where

- \( F \) = downward-directed flux at any depth, \( z \),
- \( F_0 \) = downward-directed flux at the surface, and
- \( b \) = extinction coefficient.

Figure 6 (taken from Manz (1974)) shows the measured flux profiles for three wavelength intervals; 0.42 μm < \( \lambda \) < 0.72 μm, 0.72 μm < \( \lambda \) < 0.97 μm and 0.42 μm < \( \lambda \) < 0.97 μm obtained from a natural snow sample (density = 0.28 gm/cc). The extinction coefficients for the three intervals were; 0.402 cm\(^{-1}\), 0.455 cm\(^{-1}\) and 0.421 cm\(^{-1}\) respectively.

From the data given in the figure it may be noted:

1. All flux profiles follow reasonably closely the exponential function (Equation 2).

2. A larger percentage of the incident flux of the smaller wavelengths penetrates to a greater depth than radiation of larger wavelengths. The extinction coefficient of the shorter wavelength band is less than the coefficient for the larger wavelength band.

3. For all cases, less than 50 percent of the incident flux passes below 1 cm and less than 10 percent penetrates.
below 4 cm. Note: the wavelength interval; 0.42 \mu m < \lambda < 0.97 \mu m contains approximately 61 percent of the short wave radiation (optical air mass = 1.2).

Sensible Heat Flux ($Q_H$)

The magnitude of this term is strongly influenced by the local wind velocity and the temperature gradient above the snow surface. It is not uncommon for sensible heat to reach 50 percent of the maximum net radiation on a given day even with wind speeds below 0.5 m/sec. In evaluating this term it is necessary to measure the wind and temperature profiles above the snow surface. Meaningful measurements of these profiles are not possible unless the local terrain is relatively flat and free from obstructions such as trees and buildings. This restriction makes it extremely difficult to estimate sensible heat flux in many natural situations. One instrument has been developed which is designed to measure this term directly (Bailey, Mitchell and Beckman, 1973) but it has yet to be evaluated under cold weather conditions.

Evaporation and Condensation ($Q_E$)

The magnitude of this term is governed by the vapour pressure gradient above the snow surface as well as the wind velocity gradients. During days in which melt does not occur the net amount of evaporation over a 24-hour period is generally negligible (see Figure 3). Although significant evaporation can occur around solar noon this flux is usually counteracted by equally significant condensation in the early evening. As melt progresses a small net evaporation can be measured (0.1 mm/day) and once the pack reaches a depth of less than 5 cm evaporation rates of 0.3 mm/hr have been measured. Figure 1 shows a day in which evaporation was at a maximum.

Melt Water ($Q_M$)

This term can be measured with reasonable accuracy using a properly designed snow lysimeter. On the prairies it is significant only during the last 3 or 4 days of the melt period. A great many investigations of melt water through snow have been undertaken. Recently Colbeck (1974) has developed a reasonably simple analysis of the movement of melt water through a snow pack on the assumption that the gravity force has a major influence on water movement for an isothermal snow pack. He shows a good comparison of the results of this analysis with measurements.

Ground Heat Flux ($Q_G$)

This term has a range of from 0 - 3 watts/m² at the Bad Lake Watershed and is generally insignificant in energy balance considerations. Because of its small magnitude this flux is not included in Figures 1 and
3 although it is included in the summation term of Figures 2 and 4. If an energy balance is conducted over a period of several days the cumulative affect of this flux may become important since it does not normally change direction over a 24-hour period.

**Internal Energy of the Snow Pack \( \frac{dU}{dt} \)**

For snow packs having a depth of 50 cm or less this term is a significant factor in the overall energy budget during the melt period (Figure 2) but is of less significance when melt does not occur (Figure 4). The internal energy consists of a component for the solid, liquid and vapour phases of the snow and has the form:

\[
U = \ell \left( \rho_i C_{vi} + \rho_\ell C_{v\ell} + \rho_v C_{vv} \right) T_m \tag{3}
\]

In the above expression

- \( \ell \) = the snow depth,
- \( \rho \) = the density (mass per unit volume of snow),
- \( C_v \) = the heat capacity, and
- \( T_m \) = the mean snow temperature.

The subscript \( i \) refers to ice, \( \ell \) to liquid water and \( v \) to vapour.

During non-melt days the liquid density is zero and the evaluation of internal energy involves a measurement of snow depth, mean temperature and snow density; all of which can be made with reasonable accuracy. On days when liquid water is present in the pack \( \rho_\ell \) must be measured and this we have found extremely difficult to do with a high degree of accuracy hence the large difference between the curves in Figures 2 and 4. This term causes a great deal of trouble when attempts are made to model the snow melt process. In the early part of the melt season runoff is commonly produced in the afternoon but the pack cools at night and freezes. A significant amount of energy is required the next day to bring the pack to the point where melt water is again produced. This freeze-thaw process has proven extremely difficult to model. In particular it is difficult to predict the time at which melt water will first appear on any given day. An attempt is currently being made to develop a model for the Prairies which will overcome this problem.

Dybvig (1974) found that the temperature regime within a snowpack during "non-melt" periods and hence internal energy changes could be predicted using a simple conduction model based on solution of the unsteady state energy diffusion equation. Figure 7 shows a typical set of results obtained with the model which demonstrate that the predicted and measured
temperatures are in close agreement. Dybvig tested the sensitivity of the model to different inputs; namely, net radiation and ground heat flux and air and ground temperatures. His findings indicated little improvement in the predictability of the model from using flux values over that obtained using temperature data. Table 1 gives the correlation matrix between air, ground and snow temperatures.

Table 1

<table>
<thead>
<tr>
<th>Temperatures in the Snowpack</th>
<th>TOP</th>
<th>MIDDLE</th>
<th>BOTTOM</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air Temperature</td>
<td>.987</td>
<td>885</td>
<td>.59</td>
</tr>
<tr>
<td>Ground Temperature</td>
<td>.629</td>
<td>.917</td>
<td>.968</td>
</tr>
</tbody>
</table>

Air Temperature taken at 20 cm - above ground  
Ground Temperature taken at 4.8 cm - below ground

These correlation coefficients indicate that the variation in the air and ground temperatures explain 97.4 percent and 93.7 percent of the possible variation in the temperatures at the top and bottom of the pack. This model is currently being extended to the melt period by incorporating the analysis developed by Colbeck (1974).

References


Figure 1. Energy Fluxes for April 9, 1974, Bad Lake Watershed
Figure 2. Comparison of Energy Fluxes with Changes in Internal Energy
April 9, 1974, Bad Lake Watershed
Figure 3. Energy Fluxes for March 31, 1974, Bad Lake Watershed
Figure 4. Comparison of Energy Fluxes with Changes in Internal Energy
March 31, 1974, Bad Lake Watershed
Figure 6. Broad Band Flux Profiles