A Study of Ground Heat Flux and Associated Changes in Snowpack Water Equivalent and Soil Moisture

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ABSTRACT

A study was conducted to determine if the measured ground heat flux was associated with changes in snowpack water equivalent (SWE), particularly losses in the basal snow layer, and an increase in soil moisture in the vadose zone. Results suggested that the positive ground heat flux was not sufficient in magnitude to cause any measurable loss in snowpack water equivalent and essentially no change in soil moisture conditions in the vadose zone during the period of study.

INTRODUCTION

Snow commonly represents a significant proportion of the annual precipitation received by watersheds in many parts of the world. While considerable research efforts have been directed towards understanding the relationship between snowpack melting and meltwater infiltration into soil during spring melt events (Granger et al., 1984, Marsh and Woo, 1985, Marsh, 1988), few studies have examined the extent of these interactions during the colder winter months. Soil temperature data obtained during winter months in temperate climates often reveal that the relatively warmer soil underlying a snowpack typically loses heat throughout the winter. The ground heat flux to the base of the snowpack is a component of the total snowpack energy budget, which, while usually considered as negligible during short time periods (1 week) can be potentially significant over the duration of the winter snowpack (Male and Gray, 1981). Theoretically a sufficiently high ground heat flux can slowly melt the base of the snowpack, thereby releasing meltwater to the vadose zone. Recent research conducted in the Turkey Lakes Watershed (Sault Ste. Marie, Ontario) (English et al.,1986, English et al.,1987) found that this process was likely responsible for a significant loss of SWE from the snowpack during the winter months.

The assumption that water lost from the base of the snowpack can infiltrate the soil in the vadose zone is supported by the fact that relatively impermeable saturated concrete frost rarely forms in forest soils overlaid by thick snowpacks. Studies (Garstka, 1944, Bay et al., 1952) have shown that if snowpacks begin forming late in the fall and endure throughout the winter with minimal depth losses due to melt events frost formation in the upper soil horizons is either limited to a few centimeters or non-existent. This is largely due to the ability of thick snowpacks to effectively insulate the underlying soil.

The purpose of this paper is to determine if any SWE changes within the snowpack, particularly losses in the basal snow layer, are associated with increases in soil moisture in the vadose zone.

MATERIALS AND METHODOLOGY

The following study was conducted in a small (4.1 ha.) headwater basin (Harp 4-21) located in the Canadian Shield near Huntsville, Ontario (Figure 1). In order to resolve the objective of this study the two relevant hydrologic components, the snowpack and the vadose zone, were instrumented and sampled.

Snowpack

The accurate monitoring of SWE changes in individual snow layers through time was achieved with the use of the snow time profile technique (Adams and Barr, 1974). A brief description of this technique is as follows: prior to the first snowfall event two rows of paired stakes (2 x 2") are pounded into the ground. Each pair of stakes is 2 m. apart, with 1 m. between successive stake pairs. Following a snowfall event threads are tied between each pair of stakes, thereby forming a date line for each snow layer. Based on the assumption that each snow layer is continuous throughout the site with respect to depth and density, snow pits are excavated between each pair of stakes as the winter proceeds (Adams and Barr, 1974). During the excavation of each snow pit the threads delineating each snow layer are exposed, and measurements of density and depth are taken for each snow layer. The SWE of each snow layer can then be calculated each time a snow pit is excavated between pairs of stakes.

In the Harp 4-21 basin a snow time profile site was established in late November, 1989. The site, consisting of ten pairs of stakes, was located in a lower slope area near the stream (Figure 1). The snow time profile site was located in this area for two reasons: firstly, it was located under a relatively uniform canopy of deciduous trees (<u>Acer saccharum</u>, <u>Betula alleghaniensis</u>), and secondly it was adjacent to the site where soil moisture conditions were continuously monitored. The first consideration will provide some validity to the assumption of homogeneous snow depth accumulation throughout the time profile site.

Snow pit excavations were made five times during the course of this study: Dec. 6, Jan. 8, Jan. 18, Feb. 5, and Feb. 20. During each excavation six measurements of depth and density were taken for each snow layer in the snowpack. The six depth measurements were taken across the face of the excavation within each snow layer and then averaged for a mean snow layer depth. Similarly a mean snow layer density was derived from six volumetric cores taken across the face of the excavation from each snow layer. Snow densities were calculated by weighing the snow contained in each volumetric core on a small field scale (+/- 1 g). In order to test the accuracy of the field scale several

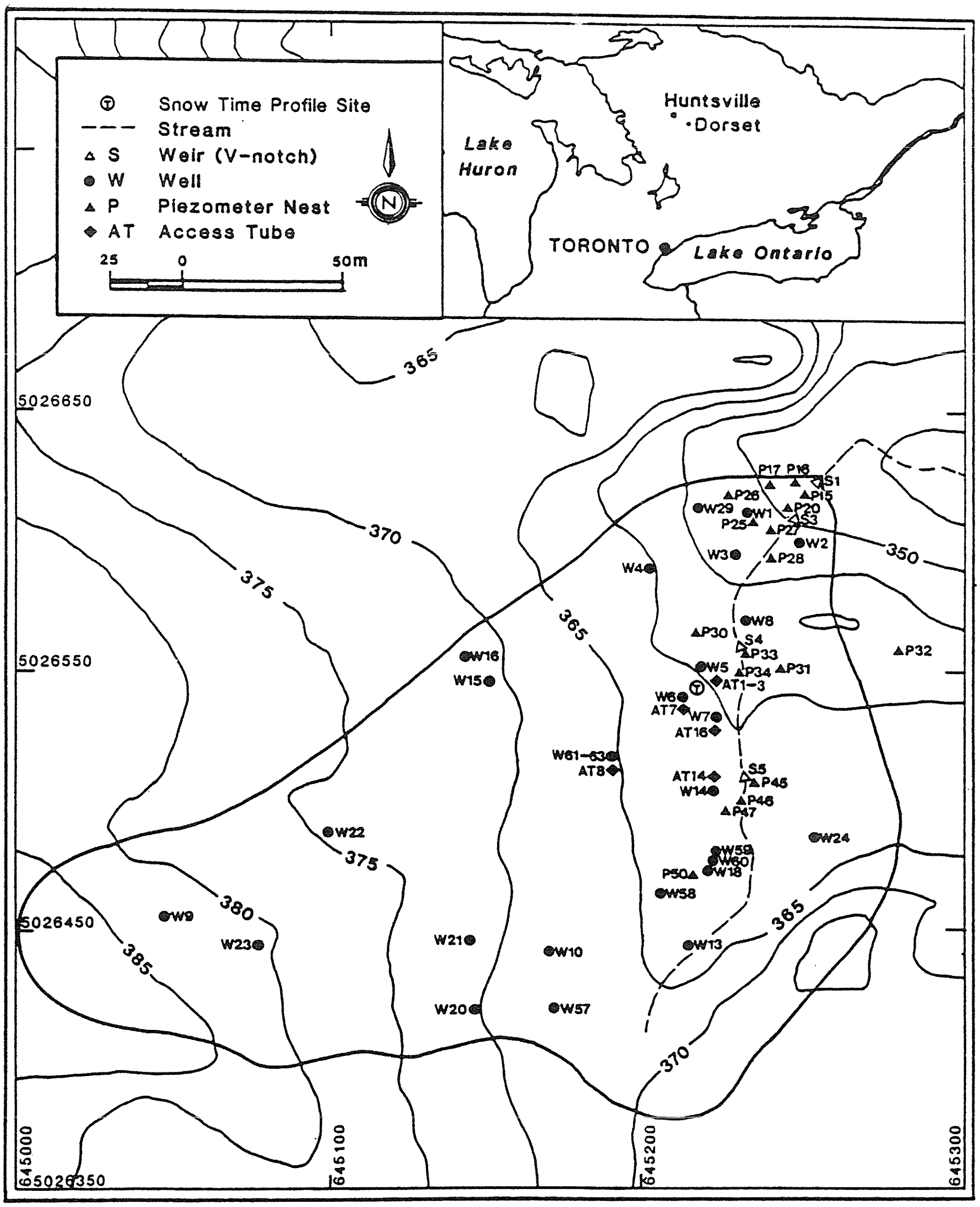

Figure 1. Instrumentation and Topography of the Harp 4-21 Experimental basin

of the field-weighed cores were periodically placed in plastic ziplock bags and taken back to the lab where the snow was melted and the volume/mass (1 ml = 1 g at 20°C) measured. The value obtained in the lab was then compared to the field result to ensure the field scale was accurate.

Vadose Zone

At a location adjacent to the snow time profile site soil temperature and moisture conditions were ascertained from an instrumented soil pit. Soil temperatures were obtained using temperature probes (Campbell Scientific model 107), while soil moisture conditions were obtained from gypsum electrical resistance blocks (Delmhorst Model 223).

Prior to installation of the temperature probes a laboratory calibration was conducted. The temperature probes and a thermometer were placed in an ice water bath which was warmed from 0 °C to 30 °C. Simultaneous temperature readings of the probes and water bath indicated a good agreement (+/- 0.5 °C). This error is close to the claimed factory (Campbell Scientific) error of +/- 0.2 °C.

Moisture conditions from the gypsum blocks are expressed as a soil water potential (bars), which is derived from a Campbell Scientific calibration of electrical resistance and bars of tension. In this study bars of tension were converted to a negative pressure head $-\psi$ (cm of water) using the equation:

$$-\psi = \text{bars} \times 10^3 \times 1.0227 \qquad (1)$$

Based on the relationship (characteristic curve) between soil water potential and volumetric water content (Figure 2) increases or decreases in soil water potential can be used to infer actual drying and wetting of the soil. Actual estimates of soil water content would be impossible without carefully derived soil characteristic curves (wetting and drying cycles) for each of the soil types in the instrumented pit.

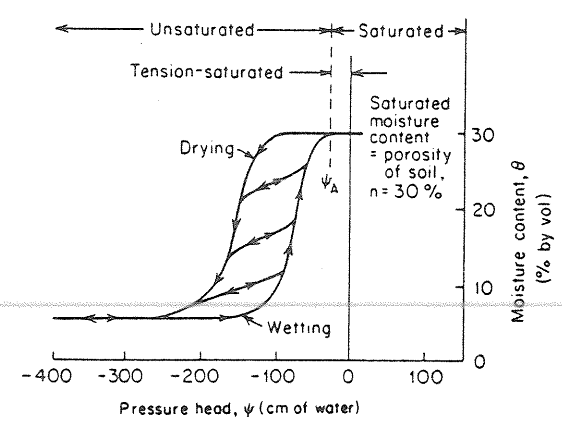

Figure 2. Soil Water Potential/Water Content Characteristic Curve
Source: (Freeze and Cherry, 1979)

Three temperature probes and four gypsum soil moisture blocks were installed in the undisturbed upslope face of an excavated soil pit (Table 1). After the instruments were installed the soil pit was carefully backfilled and the wires were run through a buried PVC pipe to an insulated box containing a Campbell Scientific 21x MICROLOGGER. The micrologger was programmed to measure the output of the temperature probes and gypsum blocks every hour and record a daily average.

Table 1: Installed Probe Depths (cm) Below Soil Surface

Soil Horizon		Temperature Probes	Gypsum Blocks
Lfh	(0 - 5 cm)		
Ah	(5 - 15 cm)	15	15
Bhf	(15 - 45 cm)	40	44
Bc	(45 - 70 cm)		70
C	(70 -)	85	85

An average ground heat flux was calculated (2) by multiplying the observed temperature gradient, determined from temperature data from the instrumented soil pit, by a soil thermal conductivity value typical of unfrozen sandy soils with 20% water content (Farouki,1986). The soil thermal conductivity value at 20% water content is essentially the same for a 40 % water content (Farouki, 1986). The soil in the instrumented pit was likely within this water content range considering that the water table was typically from 1.5 to 2.0 m below the ground surface. Based on (2) the value for thermal conductivity essentially determines the magnitude of the ground heat flux.

$$Q_f = K_f\ i \qquad (2)$$

where:

Q_f = ground heat flux (cal/cm^2/sec)

K_f = assumed average thermal conductivity (4.2 x 10^{-3} cal/cm/sec)

i = temperature gradient

RESULTS

Snowpack

In the Harp 4-21 basin snowpack accumulation began in late November and continued throughout the study period. Results of the five snow pit excavations for each of the snow layers are presented in Figure 3. There was clearly no observed loss of SWE in the basal snow layer (layer 1), as depicted by the nearly straight line, throughout the duration of this study. While the actual snow depth of layer 1 was reduced from 12.1 to 7.1 cm by the end of February there was no measured loss in SWE since the snow density increased from 0.17 to 0.32 g/cm^3 over the study period.

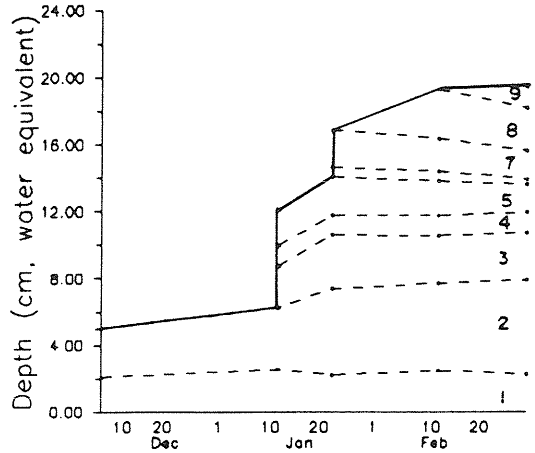

Figure 3. Snowpack water equivalent accumulation, Harp 4-21 basin

Several of the middle snowpack layers, i.e. layers 2 and 3, actually increased in SWE through time as a result of rainfall additions during periods of mild weather in January (Figure 4). Layer 4, a thick ice lens formed at the snowpack surface by a freezing rain event on January 1, 1990, was very difficult to measure for density. On the January 8, 1990 sampling several rectangular sections of known dimensions (x,y,z), hence known volume, were cut out of the ice lens and weighed in order to obtain a density value. The result was a density of 0.60 g/cm^3 which was assumed to be constant for the remainder of the study. This density value was also applied to a thinner ice layer referred to as layer 6.

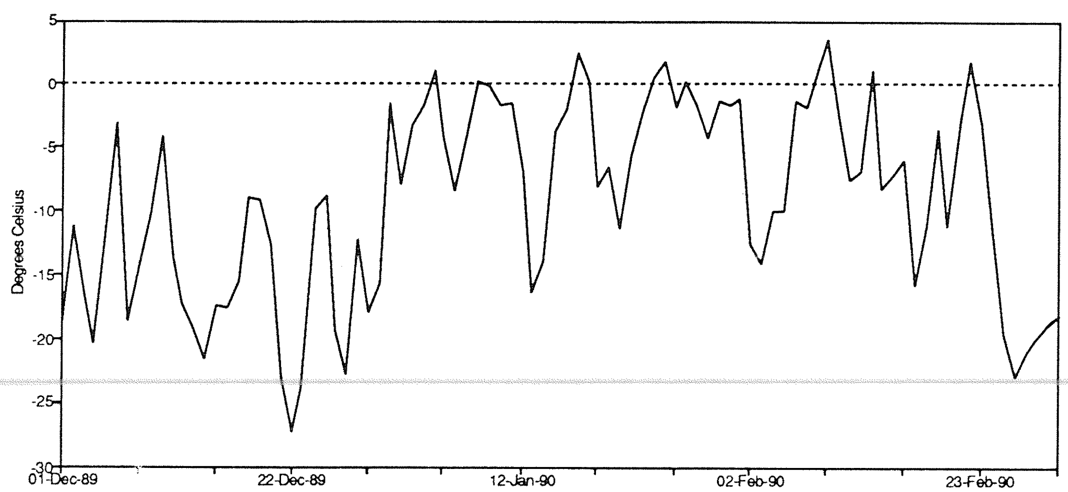

Figure 4. Daily Mean Temperature, Harp Lake

Source: (Ont. Min. of Env., 1990)

Vadose Zone

Figure 5 presents soil temperature data from the instrumented soil pit. The first, and perhaps most important observation pertains to the objective of this study: a positive ground heat flux existed throughout the study period. While each of three temperature probes indicated a gradual decrease in soil temperature during the study period, the temperature gradient remained nearly constant. For example, at the beginning of December there was a 2.4 Celsius degree difference between the 15 and 85 cm temperature probe. This difference in temperature, over a depth of 70 cm yields a temperature gradient of 3.5×10^{-2} °C/cm. By the end of February the difference between the upper and lower temperature probes is 1.7 degrees Celsius which yields a temperature gradient of 2.4×10^{-2} °C/cm. These temperature gradients, which essentially differ very little, produce a ground heat flux (2) in the order of 1.5×10^{-4} cal/cm²/sec or 13 cal/cm²/day.

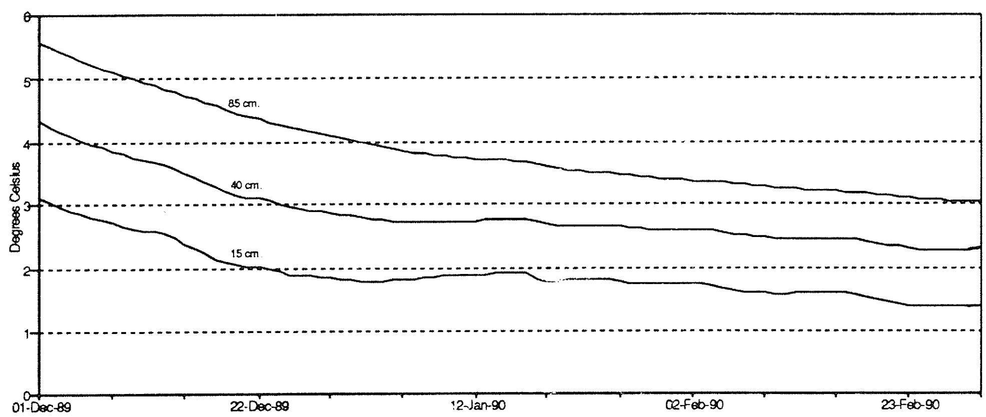

Figure 5. Daily Mean Soil Temperatures

A second important observation of the soil temperature data (Figure 5) is that the soil in the vadose zone was never subjected to freezing conditions, at least below 15 cm.. Observation made in the field during snow pit excavations confirmed the absence of frost in the Ah and Lfh soil horizons. Based on this information there was no impetus to infiltration from either melting snow or rain.

Soil water potential results from the four gypsum blocks are presented in Figure 6. Results from the gypsum blocks indicate unrealistically high negative pressure heads. Periodic depth to water table measurements from adjacent groundwater wells indicated that the water table fluctuated between 1.5 m and 2.0 m below the ground surface, hence yielding a theoretical maximum negative pressure head($-\psi$) of -200 cm. Based on this information the negative pressure head ($-\psi$) values from the gypsum blocks i.e. -395 cm. appear to be approximately 195 cm ($-\psi$) too high. While this most certainly represents a significant error in the absolute pressure head values the relative pressure head difference between the gypsum blocks is reasonably close to the actual installed depth difference.

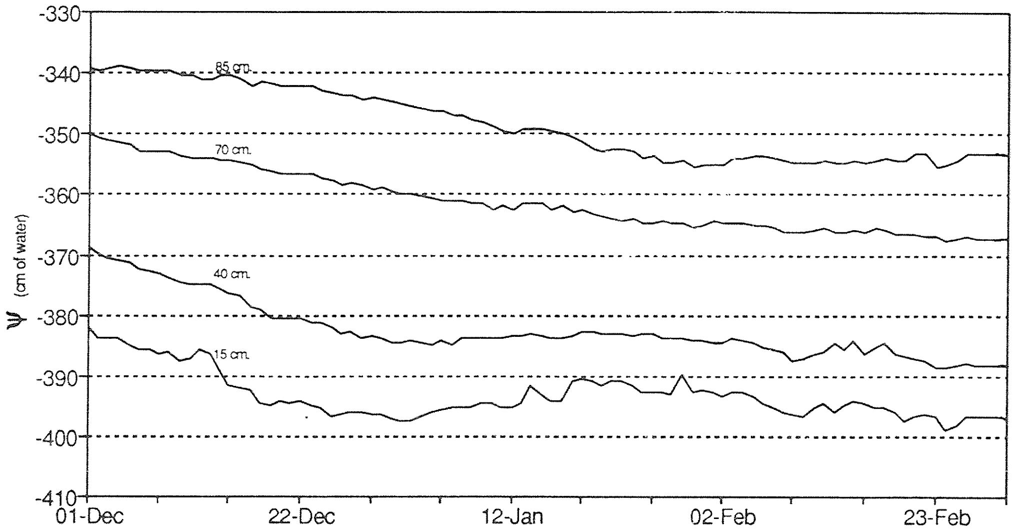

Figure 6. Soil Water Potential

Results from the 85, 70, and 40 cm gypsum blocks indicated a drying trend(gradual increase in $-\psi$) throughout the study period. Results from the 15 cm deep gypsum block indicate a drying trend in December, however, it was followed by moister conditions in January, and then a second drying trend in February. Superimposed on the general trend from the 15 cm gypsum block are two short term decreases in soil water potential (January 19, and 27) which coincide with rain on snow events.

DISCUSSION

Du´ to the periodic warming cycles and rain events throughout January and February, the month of December was the most suitable period to ascertain whether a positive ground heat flux could actually induce snow melt and subsequent increases in soil moisture. The cold conditions which prevailed throughout December (Figure 4) excluded the possibility of any snow melt from a heat flux across the top of the snowpack. Since there was no measured loss in SWE in the basal snow layer during this study, it appears that the calculated ground heat flux (13 cal/cm^2/day) was insufficient to produce any significant melting of the snowpack. If the low daily mean temperatures in December (Figure 4) had persisted throughout January and February, thereby preventing rainfall additions to the snowpack, then perhaps some SWE losses at the base of the snowpack would have been recorded.

The calculated soil heat flux in this study (13 cal/cm^2/day) is of the same order as that assumed by Wilson (1941) as an average value (20 cal/cm^2/day) during a spring snowmelt event. In terms of the overall snowpack energy budget this value is relatively small considering that the net radiation (March at 40° North latitude) entering a snowpack would be approximately 150 cal/cm^2/sec (Wilson, 1941). The low magnitude of the ground heat flux would, however, be relatively more significant in December due to the reduced net radiation component of the total snowpack energy budget.

It appears that the ground heat flux was responsible for the reduced depth and increased density of layer 1 over time. Male and Gray (1981) report that the metamorphism, or 'conditioning' of snow at the base of the snowpack during the winter can be a function of the cumulative effect of the ground heat flux. The conditioning of the snow at the base of the snowpack is an important process since it helps bring the pack to an overall state of homogeneity in terms of density prior to melt events (Male and Gray, 1981).

If there was melt at the base of the snowpack during the month of December it certainly did not infiltrate very deep since the 15 cm deep gypsum block actually indicated a drying trend in soil moisture. Under more favourable conditions, such as an early forming (October/November) thick snowpack, a high ground heat flux, and no melt across the top of the snowpack, the potential for registering an increase in soil moisture in the upper Ah horizon would be higher.

SUMMARY

There was no measurable loss of SWE in the basal snow layer during this study in the Harp 4-21 basin. The calculated average ground heat flux to the base of the snowpack was probably insufficient in magnitude to melt any significant quantities of snow at the base of the snowpack. The gypsum soil moisture blocks, which appear to yield unrealistically high soil water potentials, indicated that there was no increase in soil moisture in the vadose zone during the period of study when soil heat flux melting at the base of the snowpack would have been greatest.

Future work on this subject should consider using a better means of obtaining accurate and precise measurement of soil water content changes in the organic rich Ah and Lfh soil horizons. The technique of Time Domain Reflectometry (TDR) is well suited to this task. Further development of the time profile snow sampling technique is needed in order to validate the assumption that each snow layer is continuous, with respect to depth and density, throughout the site.

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REFERENCES

Adams, W.P. and D.R. Barr, 1974, "Techniques and Equipment for Measurement of Snowcover, Including Stratigraphy." Measurement in Physical Geography, Occasional Paper 3. Trent University

Bay, C.E., G.W. Wunnecke, and O.E. Hays, 1952, "Frost Penetration Into Soils as Influenced by Depth of Snow, Vegetative Cover, and Air Temperatures."Trans. Amer. Geophy. Union. V. 33 No. 4 pp. 541-546

English, M.C., D.S. Jeffries, N.W. Foster, R.G. Semkin, and P.W. Hazlett, 1986, "A preliminary assessment of the chemical and hydrological interaction of acidic snowmelt water with the terrestrial portion of a Canadian Shield catchment". Water, Air, and Soil Pollution V. 31 pp. 27-34

English, M.C., R.G. Semkin, D.S. Jeffries, P.W. Hazlett, and N.W. Foster, 1987, "Methodology for Investigation of Snowmelt Hydrology and Chemistry within an Undisturbed Canadian Shield Watershed". in H.G. Jones and W.J. Orville-Thomas, Seasonal Snowcovers: Physics, Chemistry, Hydrology. Reidel Publishing Co.

Farouki, Omar T., 1986, Thermal Properties of Soils. Series on Rock and Soil Mechanics V. 11. Trans Tech Publications

Freeze, R.A. and J.A. Cherry, 1979, Groundwater. Prentice Hall Inc.

Garstka, W.U., 1944, "Hydrology of a Small Watershed under Winter Conditions of Snowcover and Frozen Soil." Trans Amer. Geophy. Union. Part IV pp. 838-871

Granger R.J., D.M. Gray, and G.E Dyck, 1984, "Snowmelt Infiltration to frozen Prairie soils." Canadian Journal of Earth Sciences. v. 21 No. 5 pp. 669-677

Male, D.H. and D.M. Gray, 1981, "Snowcover Ablation and Runoff." In Handbook of Snow, Principles, Processes, Management & Use. (D.M. Gray and D.H. Male ed.) Permagon Press

Marsh, P. and M-k. Woo, 1985, "Soil Heat Flux, Wetting Front Advance and Ice Layer Growth in Cold, Dry Snow Covers." in proc. Snow Property Measurements Workshop, Lake Louise, Alta. April 1-3, 1985. Technical Memorandum 140

Marsh, P., 1988, "Soil Infiltration and Snow-melt Run-off in the Mackenzie Delta, N.W.T." in proc. 5th International Conference on Permafrost" August 2-5, 1988, Trondeim, Norway. N.H.R.I. Contribution No. 88007

Ontario Ministry of the Environment, 1990

Wilson, W.T., 1941, "An Outline of the Thermodynamics of Snow-Melt." Trans. Amer. Geophy. Union. I:182-205