Defence Science Journal, Vol 45, No 2, April 1995, pp 79-91

Effects of Capillary Discontinuities on Water Flow and Water Retention in Layered Snowcovers

Rachel Jordan

US Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, USA

ABSTRACT

The effect of capillary barriers in layered snowcovers has been examined through use of a numerical mass and energy balance model, laboratory tests and field tests. The degree of suction within the layers has been related to capillary rise and in turn to snow porosity and grain size. The relative importance of permeability and capillary tension on liquid water levels has been examined and it was concluded that capillary discontinuities play a dominant role. It has been shown both theoretically and experimentally that high-over-low suction transitions lead to interruption of water flow vertically and to horizontal movement along discontinuities. Infiltration rates predicted by the numerical model are low because of the omission of finger flow. A more realistic rendering would require a three-dimensional model or incorporate the empirical approach of Marsh and Woo[6].

1. INTRODUCTION

Water flow through snow is profoundly altered by the natural layering of a snowcover, both by textural discontinuities between porous layers and by impermeable ice lenses. Layering can induce horizontal flow along the textural boundaries and lead to flow instabilities or fingering in the underlying snow. Although the occurrence of ponding on impervious layers is intuitively obvious and has been addressed by numerous researchers, the effect of capillary barriers in snow has not received much attention. In this paper, the physics of capillarity in snow at the pore scale is reviewed, and problems and inconsistencies in the theory are presented. At the macroscale the problem is investigated theoretically using a one-dimensional mass and energy balance model of water flow through a layered snowcover. Entry pressures for the pressure-saturation curve are estimated from field observations of capillary rise for various snow types and are related to porosity and grain size. Model studies of water flow through fine-over-coarse and coarse-over-fine layerings are presented for snow of varying density and grain size. The findings of these studies are qualitatively compared with field and laboratory studies of the flow of dyed water through layered snow samples. The experimental and model investigations demonstrate that fine-over-coarse (or high-over-low suction) layerings create capillary barriers that impede the vertical movement of water.

2. BACKGROUND

In unsaturated porous media the tension or the capillary pressure P_{al} is the difference between the pressures P_a and P_l in the air and liquid water phases. The tension within the interstitial water is inversely related to the size of the filled pores and to the cumulative water content. As the water content increases, larger pores come in the picture and the overall tension decreases. Infiltrating water, therefore, backs up above a fine-over-coarse juncture until it is forced into larger pores of the same size and tension as the smallest pore in the underlying layer. After breakthrough, the re-established flow below the interface is most often in the form of fingers. The

Received 23 January 1995

presence of capillary barriers arising from fine-over-coarse layerings has been well documented in the literature on soils[1]. A similar effect has been generally described for snow by Colbeck[2], Wakahama, *et al*[3], Wankiewicz[4] and Jordan[5], but it has not been systematically or quantitatively observed. The role of capillarity in water routing has been overshadowed by the effects of large-scale inhomogeneities, such as ice lenses and by fingering, which is now generally accepted as the dominant mechanism of water infiltration into a snowcover[6]. There are still, nonetheless, many cases where capillarity is important and needs to be considered, particularly in determining the level of retained water and in predicting the formation of ice crusts. It has also been noted that wetting front instabilities are induced when the pressure gradient in the infiltrating fluid is in opposition to the flow, of which fine-over-coarse layering is a special case.

The numerical modelling of infiltration by Richard's equation, although ignores the presence of fingering, is a worthwhile step in understanding the physics of water flow through snow. A correct application of Richard's equation to layered snow requires a consistent specification of the pressure-saturation (*s-K*) and saturation hydraulic permeability (*s-p*) curves with snow characteristics. Published data on *s-p* curves in snow are scarce, perhaps because of the difficulty in making this measurement, and they do not provide sufficient information for relating the function to snow type. Because snow is a coarse material and gravitational effects play a dominant role in determining the flow rate, many numerical models, including that of this author[7], neglect the pressure term altogether. In the models[5,8,9] where it is retained, the *s-p* relationship has either not been well characterised or the effect of layered discontinuities has not been well explored.

3. THEORY

At thermodynamic equilibrium the pressure differential across an air-water interface is given by the Laplace equation:

$$P_{al} = P_a - P_l = \frac{2\,\sigma_{al}}{\gamma_{al}} \tag{1}$$

where σ_{al} and γ_{al} are the surface tension and mean radius of curvature of the interface, respectively. Although the pore shape and packing structure cannot be known for randomly distributed media, an approximate relationship between capillary pressure and pore size can be obtained by assuming a cylindrical geometry. By application of (1), the pressure drop across the meniscus in a tube of radius γ_p (or of a pore of size γ_p) is given by:

$$P_{al} = \frac{2\,\sigma_{al}\cos\theta}{\gamma_p} \tag{2}$$

where θ is the contact angle between the tube wall and the meniscus. This microscopic concept can, in theory, be extended to the macroscale by considering a cumulative distribution of filled pores, in which the capillary pressure is determined by the larger pores in the distribution. While, in practice, empirical functions are used for describing the *s-p* curve, it is important to consider the fundamental microscopic relationships among pore size, tension and water content when applying these equations. Suggested for use here is a modified form of the Van Genuchten function[10]:

$$x_s = \left\{ 1 + \left(\frac{P_{al}}{{}^{x}P_e} \right)^n \right\}^{-m} \tag{3}$$

where x_{P_e}, m and n are fitting parameters, m is set to $(1-1/n)$, and the superscript x becomes D or W for the drying or wetting branch, respectively. The parameter x_{P_e} relates to the air-entrance and air-exit pressure on the drying and wetting branches and D_{P_e} is approximately 1.5 to 2.0 times W_{P_e}. Customarily this function is scaled to an effective or mobile saturation s = (*s-sl*) / (1-*sl*) instead of *s* as shown in (3). This form of the Van Genuchten equation avoids the problem of P_{al}, incorrectly extrapolating to infinity at the immobile saturation limit *sl* and permits the curve to extend into the low-moisture region where phase change controls the water content. In the application of the expression in (3) to numerical analysis, water flow is limited to the region where *s* exceeds *sl* and the hydraulic permeability is scaled to s^*, as is the usual practice.

Based on the few published *s-p* curves for snow, a value of 3.5 is suggested for *n*. The value of x_{P_e} varies widely with snow type and is critical to characterising the capillary effects in layered snowcovers. Because of the difficulty of measuring *s-p* curves for snow, an approximate function relating x_{P_e} to snow type was developed based on field observations of capillary rise *h*. The latter is related to the capillary pressure as $h = P_{al}/\rho_{lg}$, where ρ_l, is the density of water, and g is the acceleration due to gravity. The proposed formula for the wetting branch is.

$$ {}^{W}P_e = 4\sigma_{al}\left[\frac{WA + (1-W)(1-\Phi)\,1.5}{a\,\Phi D}\right] \tag{4} $$

where D is the grain size (longest dimension), A is a fitting factor, ϕ is the porosity and $W = 2\,\phi - 1$ for $\phi < 0.5$ and is 0 for $\phi \geqslant .5$. Equation (4) utilises a hydraulic radius to define the pore size, as has been done for soils, but includes an adjustment for low-density snow controlled by the weighting factor W. It is assumed that the capillary rise corresponds to the point on the s-p curve where $s = s_l$ and to the water-exit pressure D_{P_e} and water-entrance pressure W_{P_e} on the drying and wetting branches, respectively. Because this study concentrates on infiltration, the measured values of h are for the wetting branch. The finite values for the pressure at $s = s_l$ are again an important consequence of how the Van Genuchten equation was modified. A value of 0.05 is arbitrarily suggested for s_l which is somewhat lower than that usually reported in the literature[11,12]. The lower value is believed to be more realistic for the case of water imbibition, where only a fraction of the snow is wetted[13]. The factor a in formula (4) is the ratio D_{P_l}/D_{P_e} or W_{P_l}/W_{P_e} computed from (3). A provisional value of 0.6 is suggested for A, based on a comparison of measured and estimated values of capillary rise, as shown in Fig. 1. The samples represent a wide range of snow types with porosities from 0.60 to 0.97 and grain sizes from 0.48 mm to 2.49 mm. Although the designation fine-over-coarse is used here, porosity plays an important role in determining the tension of snow, and the description high-over-low suction is actually intended. Further details on the development of formula (4) will be presented in a subsequent paper.

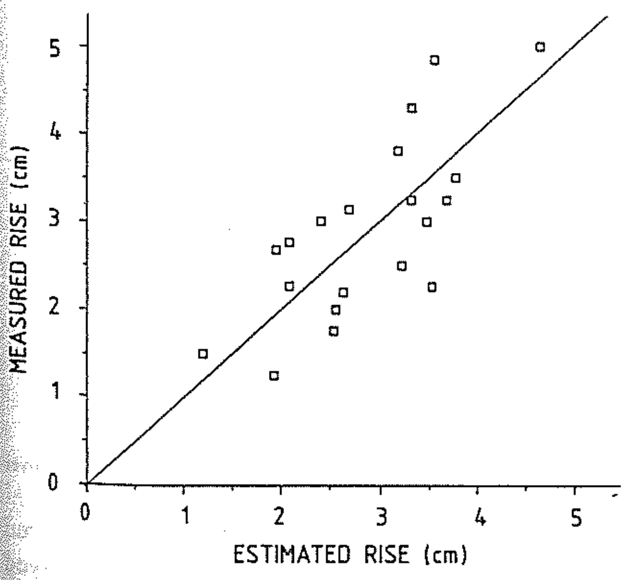


Figure 1. Measured capillary rise vs capillary rise estimated from formula (4).

The fluid flow equation is obtained from liquid water continuity, as

$$ \rho_l \Phi \frac{\partial s}{\partial t} = \frac{\rho_l \partial}{\mu_l \partial z}\left\{ K(s)\left(\frac{\partial P_{al}}{\partial z} - \rho_l g\right)\right\} + M_{il} - s\rho_l \Phi \frac{\partial V_i}{\partial z} - s\rho_l \frac{\partial \Phi}{\partial t} \tag{5} $$

where M_{il} is the melt rate, V_i is the velocity of the compacting ice matrix, Z is the vertical position relative to the snow surface, μ_l is the viscosity of water, and the change in porosity is coupled to M_{il} and V_i through the continuity equation for ice[7]. For consistency with (3), an s-K curve of the Van Genuchten[9] form is proposed:

$$ K = K_s \sqrt{{}^{x}s^*}\left[1 - \{1 - ({}^{x}s^*)^{1/m}\}^m\right]^2 \tag{6} $$

where K_s is the saturated hydraulic permeability, computed from the Carman-Kozeny equation as

$$ K_s = \frac{D^2 \Phi^3}{180\,(1-\Phi)^2} \tag{7} $$

In the numerical implementation of Eqn (5) for infiltration, a maximum value for P_{al} is set at the water-entrance pressure W_{pl}.

4. RESULTS

4.1 Laboratory Experiments

The theory of water flow through fine-over-coarse and coarse-over-fine junctures was tested with two controlled experiments and several field tests. In the first experiment, two-layer snow samples of both juncture types were constructed inside plastic tubes. The samples consisted of natural snow with a density around 150 kg/m^3 and the same snow compacted to a density of about 500 kg/m^3. After introduction of dyed water at the bottom by capillary action, the tube was instantly flipped over to allow the imbibed water to percolate downward. The dye was blue food colouring with an electrical conductivity of about 10^{-4} mho/cm and was judged to have a negligible effect on the freezing

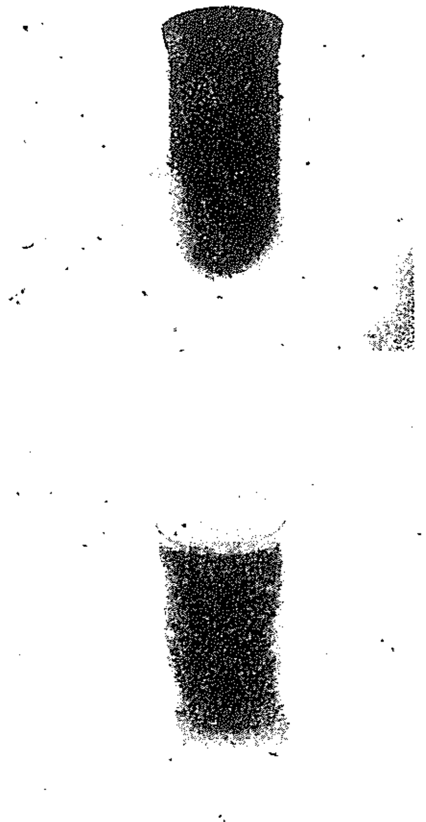


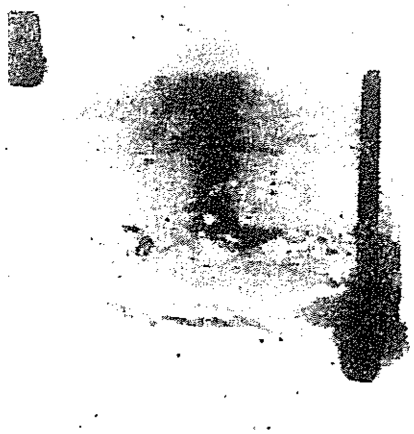
Figure 2. *Top* : Backup of imbibed dye-water above a high-low suction junction. *Bottom* : passage of dye-water through a low-high suction junction.

point. In agreement with theory, the water passed without interruption through the coarse-over-fine juncture (Fig. 2, bottom) and noticeably backed up about the fine-over-coarse juncture (Fig. 2, top). It is interesting to note that a certain amount of water remained within the fine (or lower) layer in the coarse-over-fine case. This water is held against gravity because of the fine-coarse transition between the snow at the bottom of the tube and the air.

In the second experiment, a coarse-fine-coarse layering was constructed *in situ* utilising the natural snowcover, as shown in Fig. 3. The middle and lower layers consisted of 12 cm of fine-grained snow sieved onto a base of coarser natural snow, and the upper layer consisted of 9 cm of poured coarse snow. The layers (bottom to top) were characterised by densities of 225, 425 and 347 kg/m^3 and capillary rise measurements of

3 (a)

3 (b)



3 (c)

Figure 3. Infiltration of dye-water into a layered snow cover consisting of low-high-low suction layers.

2.25, 8.75 and 4.0 cm (Table 1). Samples were cored horizontally with a 5.0 cm diameter plastic tube. Dyed ice water was sprayed onto the top snow surface somewhat behind the vertical face using a hand-pumped atomiser. The optimal irrigation rate for a circular spray area of 25 cm diameter was ~ 18 cm/h, but there was a tendency for the nozzle to freeze up, and the actual rate was much slower. As is typical with irrigation studies of this type, water initially built up at the surface until it reached a saturation threshold. It then flowed very rapidly through the top layer and reached the middle layer in a matter of seconds (Fig. 3a). Water passed through the coarse-over-fine juncture without interruption and into the fine-grained middle layer, where it flowed at a somewhat slower rate. It reached the bottom layer in an additional 1.5-2 min (Fig. 3b). Infiltration was halted at the capillary barrier between this fine-over-coarse juncture, and the flow spread horizontally along the interface (Fig. 3c). Irrigation was continued for another half hour. Dyed water spread across the entire 45 cm face of the prepared snow block and backed up considerably, but it never penetrated the juncture. This was verified on the following morning after allowing the test area to freeze overnight. Water had eventually flowed through to the ground, where it formed a solid ice layer. The flow path, however, was through a high-suction vertical ice crust that had been used to help stabilise the wall of the test block. Snow underneath the plane of the juncture within the test block remained totally dry, with no indication of any penetrating flow fingers.

4.2 Model Simulations

The three-layer experiment described in Sec. 4.1 was modelled numerically with the one-dimensional SNTHERM model developed at the Cold Regions Research and Engineering Laboratory (CRREL) in Hanover, New Hampshire. The basic model is that described in Jordan[7], with modifications for capillary flow and use of the Clapeyron equation to predict the relationship between liquid saturation and temperature depression. Simulated s and P_{al} profiles at 5 min intervals are shown in Fig. 4. An irrigation rate of 5 cm/h was assumed, and the initial snowpack temperature was taken as -0.05 °C. Air exit values of 63, 262 and 118 P_a were used for the lower, middle and top layers, respectively, giving corresponding water entrance values of 209, 864 and 391 P_a. Grain sizes were selected so that the water entrance values computed using relations (3) and (4) approximated the suction measured by the capillary rise (Table 1). Nodal thicknesses were 5 mm except across the bottom juncture. Because of the nonlinear water saturation profile, a finer mesh of 2 and 3 mm was used about this interface. Too coarse a mesh underestimated the amount of liquid water held above the discontinuity. The simulated infiltration rate is slower than the observed rate because the model did not account for finger flow, but the essential effect of

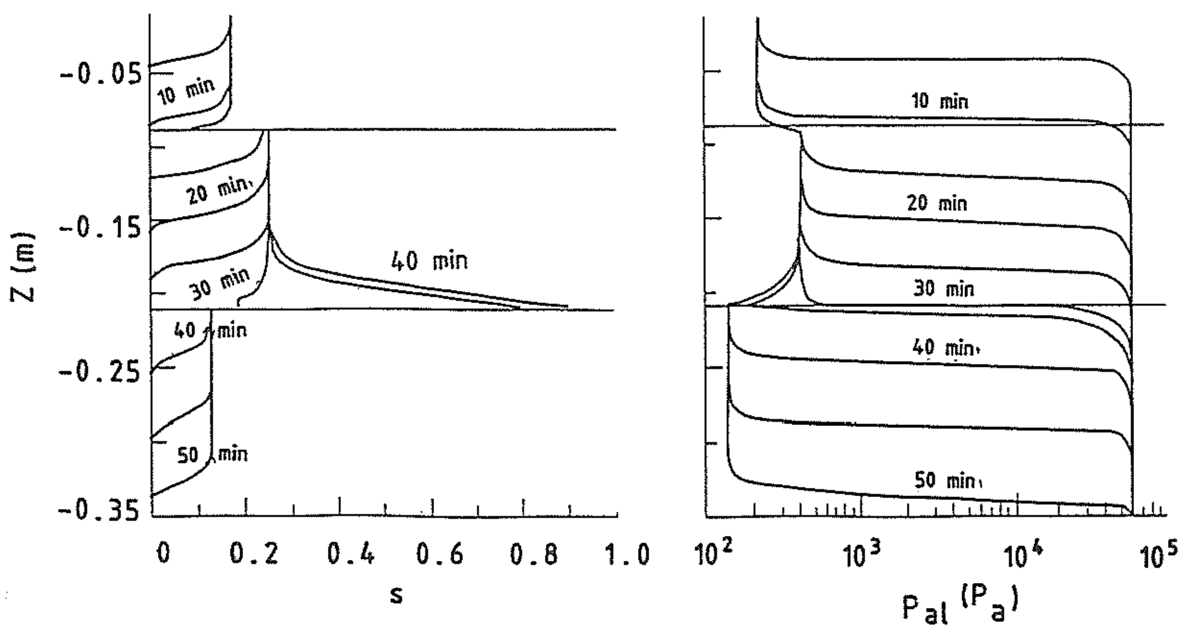


Figure 4. Simulated profiles of liquid saturation and capillary pressure at 5 min intervals for an irrigation rate of 5 cm/hr. The vertical position Z is relative to the snow surface.

Table 1. Snow characteristics, *s-p* curve parameters and saturated hydraulic permeability for the three-layer experiment

Layer	Density (kg/m^3)	Grain size D (mm)	Capillary rise h (cm)	Air-exit pressure $^{W}P_e(P_a)$	Water-entrance pressure $^{W}P_l(P_a)$	Hydraulic permeability K_s(m^{2}10^{-10})
Top	347	0.70	4.0	118	391	45.6
Middle	425	0.45	8.75	262	864	8.07
Bottom	225	0.70	2.25	63	209	193.0

capillary barriers can be noted. Once steady-state conditions have been obtained and if melt and compaction are ignored, the values of s and P_{al} must satisfy the relationship

$$u_p = \frac{\rho_l K(s)}{\mu_l}\left[\frac{\partial P_{al}}{\partial z} - \rho_l g\right] \tag{8}$$

where u_p is the irrigation rate. At the top of each layer the pressure gradient is negligible and s approaches the expected value for gravitational flow, dependent on the saturated hydraulic permeability for that snow type. Across the junctures, P_{al} is continuous while there is a discontinuity in s. As the coarse-fine and fine-coarse junctures are approached, there is a decrease and increase in s, respectively, to maintain continuity in P_{al}. The points on the s-p curve for the lower and middle layers during steady-state flow (40 and 50 min) are shown in Fig. 5 and illustrate the buildup in s above the juncture.

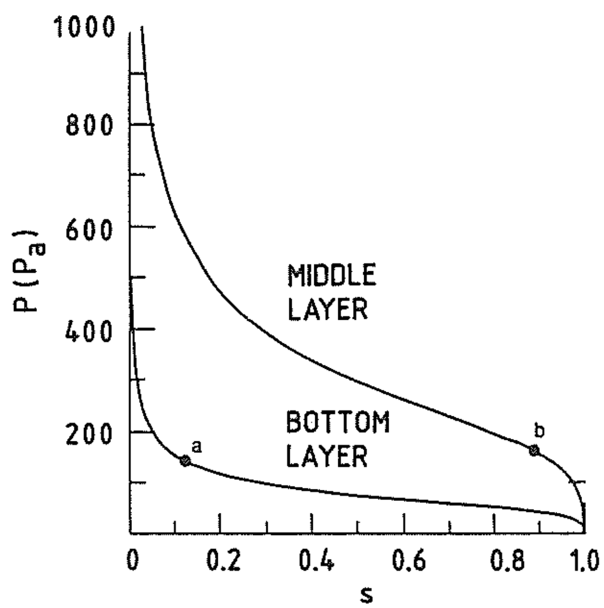


Figure 5. Pressure saturation curves for the bottom and middle layers in the three-layer experiment, where the water exit. values are 63 and 262 P_a. The points *a* and *b* are the *s* and P_{al} for steady-state flow.

It is of interest to consider the relative importance of permeability and capillary barriers in determining the steady-state water content of layered snowcovers. The relationships are most easily grasped for the low saturation regime, where relations (3) and (6) can be simplified using a first-order Taylor's expansion and become, respectively,

$$x_s = \left(\frac{^{x}P_e}{P_{al}}\right)^{nm} \tag{9}$$

$$K = K_s m^2 (\,^{x}s^*)^{2/m+0.5} \tag{10}$$

when n is taken as 3.5, the power on s^* in Eqn (10) becomes 3.3, which is close to the value of 3 suggested by Colbeck and Anderson[14]. In the gravitational limit the relative water content is governed by K_s and s^* varies approximately as $1/(K_s m^2)^{-(2/m+0.5)}$ or as 1.995 $K_s^{-0.30}$ when n is 3.5. The values of K_s for the bottom through the top layers are 193.0, 8.07, and 45.6 m^2.10^{-10}, giving predicted ratios of 2.59 and 0.60 when comparing s^* for the bottom two layers and top two layers, respectively. This compares favourably with the numerically modelled ratios of 2.44 and 0.62. At low saturations, the discontinuity in s across a juncture varies as $(^{x}P_e)^{nm}$ as a result of the required continuity in P_{al} or as $(^{x}P_e)^{2.5}$ when n is 3.5. Although the low-saturation approximation is not appropriate for the present case, where s is 0.89 above the bottom juncture, it can be seen from the relative powers of -0.30 and 2.5 on K_s and $^{x}P_e$, respectively, that capillary barriers lead to the larger discontinuity in s.

4.3 Field Tests

During January and February of 1994, field tests were conducted at a site near CRREL (72°15′45″W, 43°45′30″N) on the flow of dyed fluids through natural snowcovers. Flow experiments were documented with photographs, snow characterisation of the various layers and measurements of capillary rise. Grain sizes were obtained from formvar replicas and represent a mean of the largest dimension of about 150-250 individual grains. A site was also maintained on the CRREL grounds for validating the SNTHERM mass and energy balance model. Standard meteorological measurements and temperature profiles were recorded every 10 min at this site, and snow characterisation was done on a weekly basis or following snowstorms. The flow site was in a clearing surrounded by coniferous trees and had a deeper snowcover than the more exposed site on the CRREL grounds. The stratigraphy between the two

Table 2. Precipitation events during 1 January 1994 – 10 March 1994

(Measurements of equivalent mass of water and snowfall were made on the CRREL grounds by the USA TECOM Hanover Meteorological Team. Because snowfall was recorded once daily at around 0800 hrs, the start and end times of the events as well as the precipitation type are from summaries prepared at Lebanon Airport, 6 miles to the south. Breakdowns of equivalent mass of water and snowfall within storms are computed in proportion to automatically recorded values at a site near South Royalton, Vermont, 16 miles from CRREL)

Layer	Start	End	Type	Water equiv mass (mm)	Snowfall (cm)
	1 January 2300	2 January 0100	–	1.0	T*
A	3 January 1300	3 January 2110	Snow	T	T
	4 January 0400	5 January 0130	Snow	10.3	13.1
	5 January 1130	5 January 1430	Snow	2.3	0.6
	5 January 1600	5 January 1700	Snow	1.1	0.3
B	6 January 1900	7 January 0700	Snow	3.0	4.0
	7 January 1500	8 January 1300	Snow	12.6	18.0
	12 January 1500	12 January 1900	Snow	T	1.0
C	14 January 1100	15 January 0100	Snow	3.4	4.5
D	17 January 0630	18 January 0800	Snow	26.4	41.0
	22 January 0800	22 January 1500	Snow	T	T
	23 January 1100	24 January 0800	Snow	0.2	–
	26 January 0200	26 January 0800	Snow	0.3	1.0
E	27 January 2200	28 January 0700	Snow	–	–
	28 January 0700	28 January 0900	Snow/sleet	28.1	8.0
	28 January 0900	28 January 1400	Frz rain	–	–
	28 January 1400	28 January 1700	Rain	2.1	–
F	8 February 2200	9 February 2300	Snow	7.3	20.5
G	13 February 0200	13 February 1600	Snow	0.4	1.0
H	15 February 2300	16 February 0900	Snow	T	1.0
I	23 February 1100	24 February 0300	Snow	3.7	3.0
	24 February 0800	24 February 1300	Sleet	9.8	4.0
	24 February 1300	24 February 2000	Snow	–	–
J	3 March 0500	4 March 1100	Snow	14.8	24.0
K	7 March 2000	8 March 0630	Rain	5.0	–
L	9 March 2100	10 March 0300	Snow	5.3	9.0
	10 March 0300	10 March 1330	Rain	17.7	–

*T signifies that snowfall is only trace.

sites was similar, however, and simulations run on the meteorological data from the CRREL site were used to provide estimates of the historical occurrence of liquid water in the pack.

The winter of 1994 was unusually snowy for the New England region. Between 3 January and 10 March 1994, there were about 17 precipitation events totalling 154 cm in equivalent mass of water, as summarised in Table 2.

The new snow varied widely as to density, crystal size and type and therefore afforded a very heterogeneous snowcover with numerous snow layers and ice crusts. Several of the flow experiments were conducted soon after snowfalls. Typically the most distinct examples of horizontal spreading along discontinuities occurred near the surface at junctures between new and older snow and within ice crusts. Flow transitions lower in the pack were less distinct and involved only slight jogs in the flow path. This was judged to be the result of snow metamorphism, leading to a more homogeneous snowpack over time. Surface effects, such as wind packing, melt-freeze crusts and rain glazes, increased the suction of the exposed snow surface and may have accentuated capillary discontinuities between new and old snow layers. With the exception of one case, all field observations agreed with the proposed theory of horizontal flow along fine-coarse boundaries. Following are summaries of the more interesting features of the field tests.

4.3.1. Case 1: 27 January 1994, 1300-1500 hrs

The snowcover of 27 January 1994 was 51 cm deep and contained layers from several snowfalls, beginning with that of 4-5 January 1994, which fell onto bare ground. Air temperatures for this period were below normal (Fig. 6), and the snow remained cold and dry, except for 14 January 1994 from ~ 1200 to 1900 hrs and 24 January 1994 from ~ 1500 to 1730 hrs, when the SNTHERM simulation showed melt occurring within the top 1.0-2.0 cm of the pack. The former melt period coincided with a wet snowfall of 14 January 1994 and resulted in a barely discernible crust, which was buried by the large snowstorm of 17-18 January 1994. The approximate divisions between the more substantial storms of 3-27 January 1994 are indicated in Fig. 7. Measurements of capillary rise were made at heights of 5.0, 15.5, 30.5 and 46.0 cm above the snow-soil interface and are noted along with snow characteristics in Fig. 7. Capillary rise values ranged from 2.50 to 3.75 cm, and there were no marked differences between the layers. The uncertainty in this measurement is about 0.5 cm for homogeneous snow. In instances where the snow characteristics varied across the 5 cm sampling diameter, a corresponding variation in the capillary rise was observed. Density measurements generally agreed with a more complete profile done at the CRREL site, where the profile was relatively smooth and monotonic. Flow experiments were conducted using both dyed water and dyed mineral spirits (a petroleum distillate) as fluids.

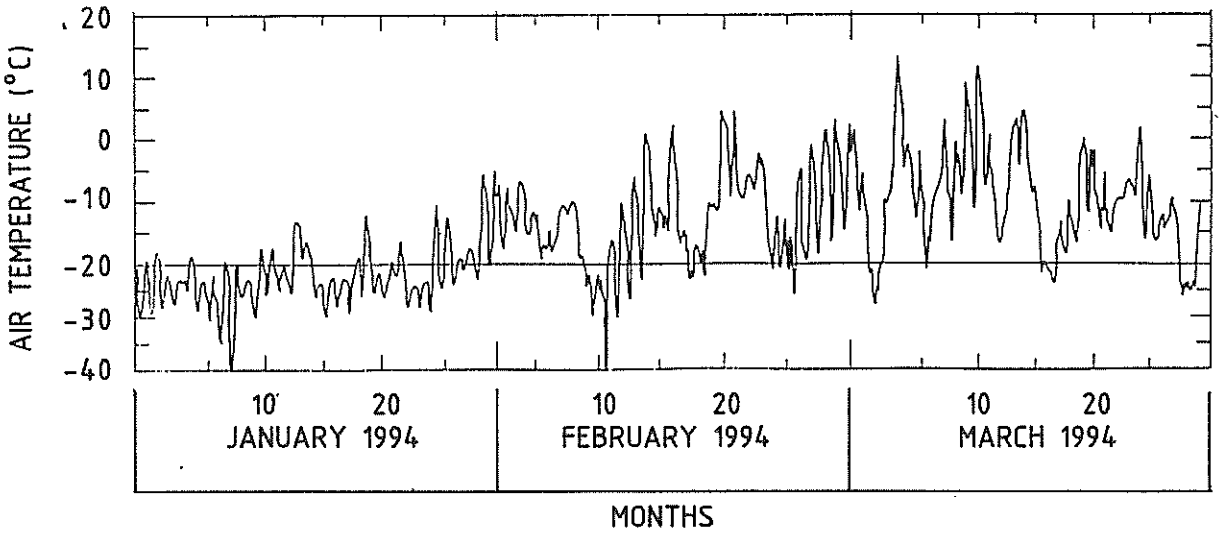


Figure 6. Plot of air temperature recorded at 2 m height at Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, for January 1994 through March 1994.

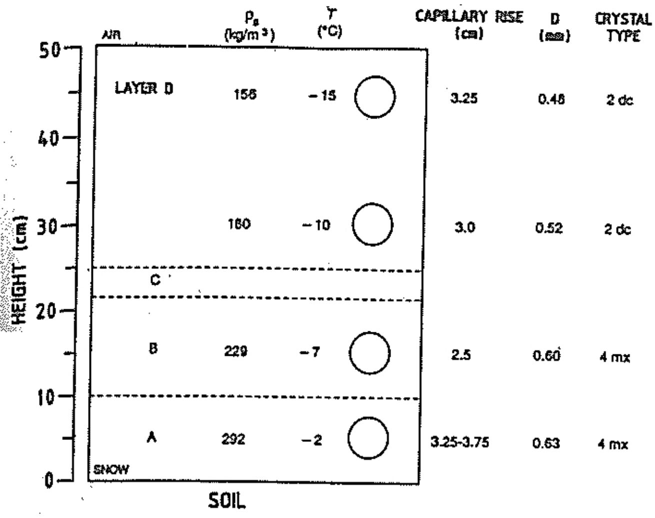


Figure 7. Snow characteristics and capillary rise measurements for flow tests of 27 January 1994, 1300-1500 hrs (Case 1) at flow site in Hanover, New Hampshire. Codes for the grain types are according to The International Classification for Seasonal Snow on the Ground, Colbeck, *et al*[15].

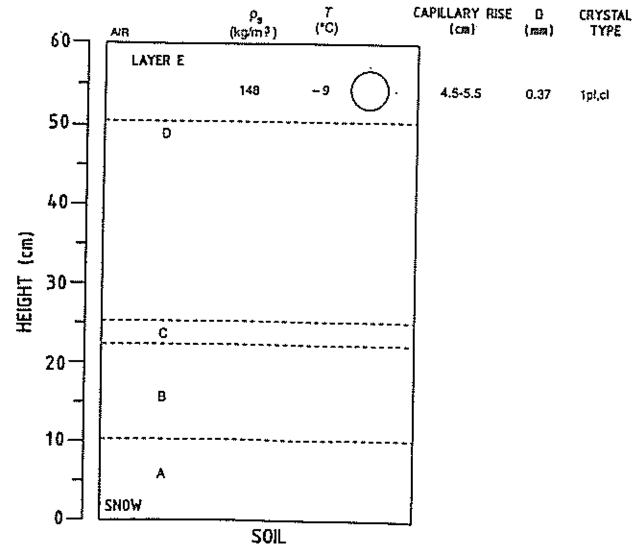


Figure 8. Snow characteristics and capillary rise measurement for flow tests on new snow of 27-28 January 1994, conducted on 28 January 1994, 0700-0800 hrs (Case 2) at flow site in Hanover, New Hampshire.

As with the three-layer laboratory experiment, the fluids were sprayed onto the snow surface with a hand-pumped atomiser. In both cases infiltration occurred rapidly without interruption, following an initial period of fluid buildup at the surface. Flow patterns of the two fluids were similar, and freezing was judged to have had a minor effect. The flow path was almost straight down except for a slight jog at about 20 cm above the soil. This may have corresponded to a fine-coarse transition between snow layers *C* and *B*, although the resolution of the characterisation was not sufficient to make a determination.

4.3.2. *Case 2: 28 January 1994, 0700-0800 hrs*

Snowfall began at 2200 hrs on the night of 27 January 1994 and turned to rain at around 0900 hrs on the following morning. With the passage of a warm frontal occlusion aloft, air temperatures rose overnight from a low of -16 °C at 1820 hrs to -9 °C at 0800 hrs and to a maximum of 4 °C at 2000 hrs on the following afternoon. Prior to the rainfall transition, two water flow tests were conducted to observe the effect of the new snow layer on the pattern. As noted in Fig. 8, the new snow was relatively dense (148 kg/m^3), fine-grained (0.37 mm), and consisted primarily of hexagonal plates, columns and sleet. It had a correspondingly high capillary rise of 4.5-5.5 cm and a high suction, as compared with a rise of 3.25 cm for the underlying layer. As predicted by theory, flow was interrupted at the new-old snow interface, and water moved horizontally along the discontinuity. After ~0900 hrs rain began to freeze on the snow surface, forming an ice crust that grew to a depth of 1.0-1.5 cm before the rain stopped late in the afternoon. The wet ice crust remained permeable and had a capillary rise of 4 cm at 1530 hrs when immersed vertically in dyed water. After the crust was placed in a freezer overnight, the rise increased to 5 cm, and there was a noticeable decrease in the imbibition rate.

4.3.3. *Case 3: 10 February 1994, 1300–1530 hrs*

On the afternoon of 9 February 1994, prior to the test, a section of ice crust about 1 sq m was removed so that the accumulating new snow would form a snow-snow junction with the underlying layer. About 20 cm of very light snow fell during this storm in the form of large dendrites, particle fragments and hexagonal plates. The snow is characterised further in the discussion of Case 4. The primary objective of this test was to assess the feasibility of using a time delay reflectometer (TDR) probe to detect a water pulse during flow infiltration. For convenience, the test was conducted near the laboratory building a few feet inwards from a walkway. The choice of location led to anomalous but interesting results in the flow pattern. Distinct horizontal flow leading to very intact dyed ice sheets 15-25 cm in diameter occurred at four levels within the snowpack. After some reflection it was concluded that the layers corresponded to wind-packed snow from the snowblower used to clear the walkway. The effect of road salt on the freezing point was also considered and discounted, since analysis of samples of the melted lenses showed extremely low salt concentrations.

4.3.4. Case 4: 11 February 1994, 0830-1000 hrs

As in Case 3, a section of the surface ice crust was removed on the afternoon on 9 February 1994, but now in an undisturbed snow area at the usual water flow site. The storm was already in progress, and an additional 13 cm of new snow fell on the exposed base from the storm of 27-28 January 1994, as shown in Fig. 9. At 0830 hrs on 10 February 1994 the density and capillary rise were measured as 27 kg/m^3 and 1-1.5 cm, respectively. Because of the fragility of the snow, it was

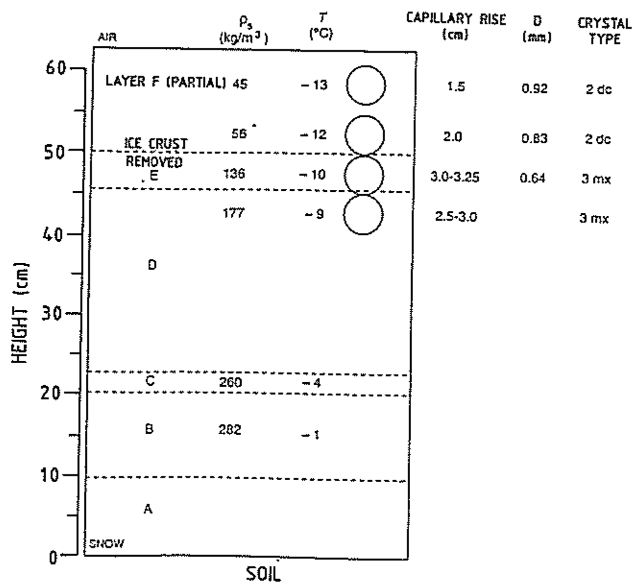


Figure 9. Snow characteristics and capillary rise measurements for flow tests of 11 February 1994, 0830-1000 hrs (Case 4) at flow site in Hanover, New Hampshire. The ice crust of 28 January 1994 was removed prior to the storm of 9 February 1994.

difficult to obtain samples, and these measurements are uncertain. An average density of 36 kg/m^3 for the storm, computed from the water equivalent mass and snowfall, may be more realistic. In computing the grain size of 0.92 mm the individual arms of the large dendrites rather than the complete particles were averaged. By the time of the test, the snow had densified to 45-56 kg/m^3 and the capillary rise had increased to 1.2-2 cm. Dyed water was sprayed on the snow surface always behind the vertical face. No water penetrated the face, suggesting an absence of significant horizontal flow. From the start of irrigation it took about 3 min for the water front to reach the new-old snow junction, the progress being observable through the semi-transparent new snow. Flow was slow because the atomiser nozzle repeatedly froze up in the cold air (-14 °C). Spraying was stopped after 23 min when ponding was observed at the soil interface. On the following morning, loose snow was brushed away from the refrozen flow area to reveal the flow pattern. In contrast to Case 3, where there were four notable ice sheets, refrozen snow was confined to the columnar section below the irrigation area. The flow pattern through the new snow was in the form of fragile fingers 0.5 cm in diameter, which broke easily during excavation. More water was retained in the older snow, and the flow fingers were thicker. Of particular interest in this case was the absence of horizontal flow at the new-old snow junction, which is consistent with a coarse-fine transition. At first this would appear to be a good example of horizontal flow not occurring at a new snow interface. Because the junction was artificially created, the microstructure may not have been realistic since there was no opportunity for surface metamorphism of the underlying layer.

4.3.5. Case 5: 15 February 1994, 1030-1230 hrs

The snow cover stratigraphy was similar to that of 11 February 1994, except that the ice crust was not removed (Fig. 10). On 13 February 1994 there was an additional 1 cm of new snow. Flow tests were conducted using both dyed water and mineral spirits. In both cases two noticeable horizontal layers formed, one within the snow layer of 9 February 1994 at 5 cm above the ice crust and another within the ice crust. After the start of irrigation the mineral spirits took ~3.0 min to reach the first juncture, 20-25 min to penetrate the ice layer and a total of ~35 min to reach the ground. A photo of the flow pattern for this test is shown in Fig. 11, indicating the distinct horizontal layering and the vertical pattern below the ice crust. Formation of the upper horizontal layer was anomalous in that capillary rise measurements above and below the junction were 1.5-2 cm and 3.25 cm, respectively, making this a coarse-fine transition. To gain further information about the microstructure, a digitised image was prepared from an aggregated snow sample containing the junction. Analysis of the image using the technique of stereology confirmed the presence of a coarse-fine junction and did not reveal any other outstanding features. Subsurface melt in the snow above the ice crust had been observed at the CRREL site on 12 February 1994, which may have led to the formation of a thin ice layer not resolvable by stereology.

It is also of interest to observe the effect of the ice crust on water flow. The refrozen area for the water flow test is shown in Fig. 12, where the loose snow has been brushed away down to the crust. The irrigation area is to the right of centre. There was a slight incline to the site, and three horizontal flow channels about 5 cm across, 11 cm apart and 0.5-1 m long were observed within the crust. Since the snow above and below the crust was powdery and had never been wetted, the high-suction ice crust clearly acted as a conduit for water flow.

4.3.6. *Case 6: 24 February 1994, 1000-1200 hrs*

A snowstorm consisting of small stellar particles, columns and needles began around 1000 hrs on 23 February 1994. It stopped around midnight and then resumed as sleet at 0800 hrs on 24 February 1994, turning back to snow at 1300 hrs before stopping at 2000 hrs. At 1530 hrs on 23 February 1994 the ice crust of 28 January 1994 and the snow above it were removed

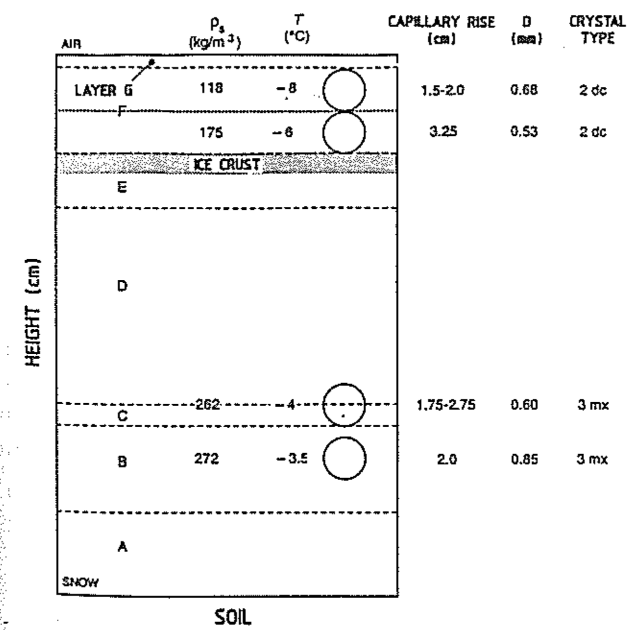


Figure 10. **Snow characteristics and capillary rise measurements for flow tests of 15 February 1994, 1030-1230 hrs (Case 5) at flow site in Hanover, New Hampshire.**

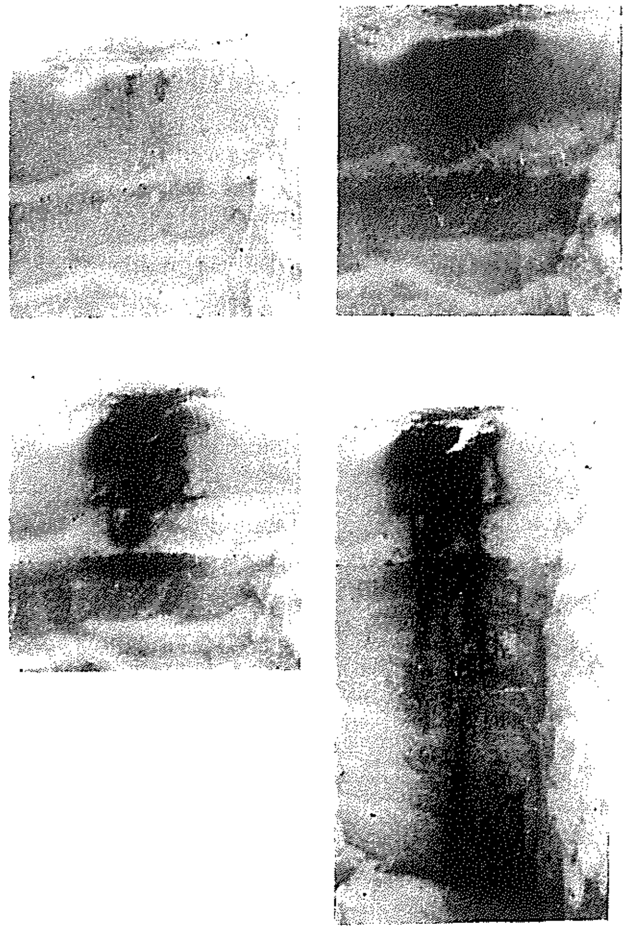


Figure 11. **Infiltration of dye-water into the natural snow cover of 15 February 1994 (Case 5).**

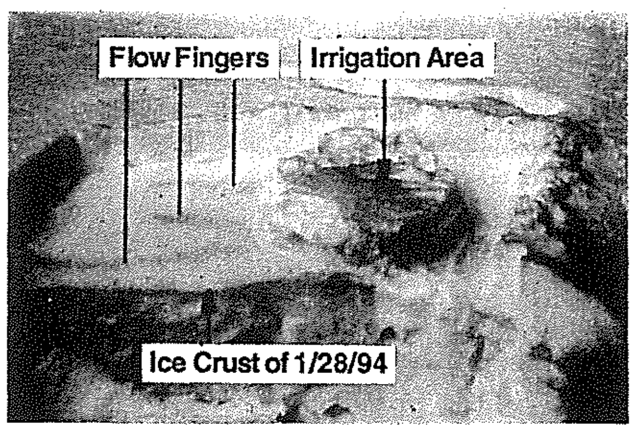


Figure 12. **Horizontal channeling of flow fingers by an ice crust. (Case 5).**

from a 1 sq m area of the test site. Between then and the time of the test, an additional 6 cm of new snow had fallen (Fig. 13). The densities of the upper and

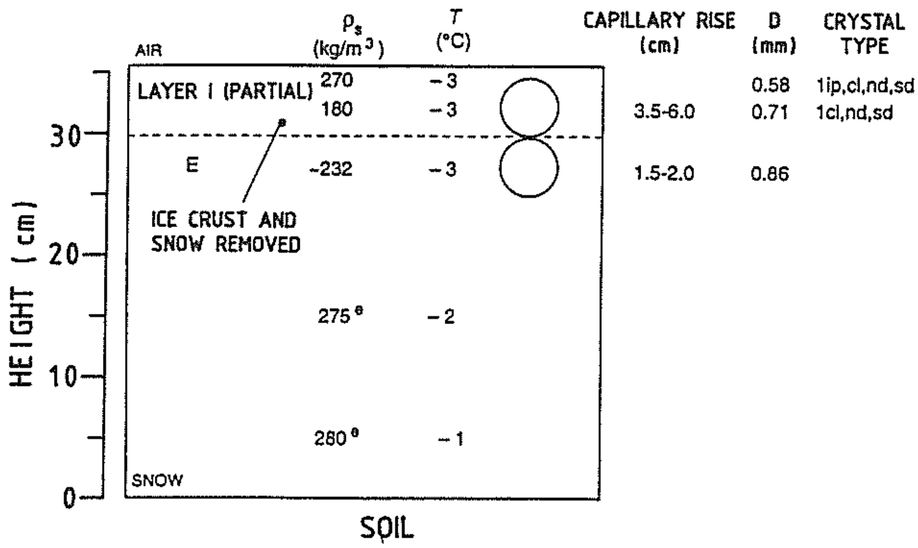


Figure 13. Snow characteristics and capillary rise measurements for flow test of 24 February 1994, 1300-1500 hrs (Case 6) at flow site in Hanover, New Hampshire. The ice crust of 28 January 1994 and the snow above it, including the earlier snow of 23 February 1994, were removed partway through the storm. The superscript *e* indicates estimated values.

lower halves of this layer were 270 and 180 kg/m^3, respectively, with the positive density gradient reflecting the occurrence of sleet in the upper portion of the snow cover. There was also a marked gradient in capillary rise for the new snow layer, ranging from 3.5 cm at the bottom to 6 cm at the top. The capillary rise of the underlying snow was 1.5-2 cm, making this a fine-coarse transition. As noted in Fig. 14, a distinct horizontal layer formed between the new and older snow, as would be predicted by theory. The flow was otherwise vertical and unimpeded.

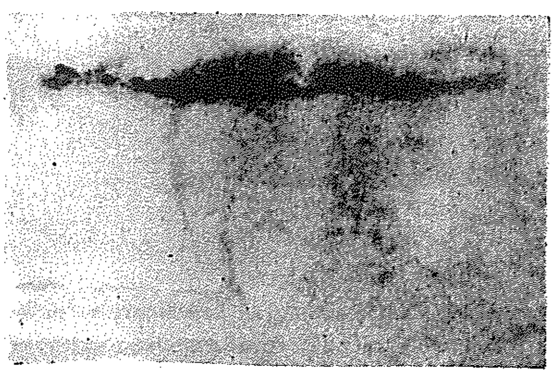


Figure 14. Infiltration of dye-water into partially altered snow cover of 24 February 1994 (Case 6).

5. DISCUSSION

Horizontal spreading was observed to occur most frequently between new and old snow layers. While it can be said conclusively that fine-over-coarse layerings result in capillary barriers, the converse of no flow occurring along coarse-over-fine boundaries is uncertain. In the laboratory tests where the junctures were formed from cored, sieved or poured snow, or in Case 4 where the surface snow was first removed, water flowed without interruption through coarse-over-fine boundaries. In the case of natural snowcovers, however, changes in the microstructure of the exposed snow surface may lead to capillary discontinuities not predicted by the bulk properties of the layer. From an examination of the relative importance of permeability and capillary tension on liquid water levels in layered snow, it was concluded that capillary discontinuities play a more dominant role. Thus permeable ice crusts tend to act as conduits for water flow, as was observed in Case 5 and in the vertical channeling of water through an ice crust used as a support in the three-layer laboratory test.

ACKNOWLEDGEMENTS

The author thanks Dr. R. Davis and Dr. P. Black for technical reviews of this report. Funding was provided by DA Project 4A762784AT42. The author very much appreciaties the assistance of Mr. Bryan Harrington and Mr. John Fiori in the collection and processing of data, and of Ms. Nancy Kent in the preparation of the graphs.

REFERENCES

1. Baker, R.S. & Hillel, D. Laboratory tests of a theory of fingering during infiltration into layered soils. *Soil Sci. Soc. Am. J.*, 1990, **54**, 20-30.
2. Colbeck, S.C. Effects of stratigraphic layers on water flow through snow. US Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 1973. Research Report 311.
3. Wakahama, G.; Nkamura, T. & Endo, Y. Infiltration of melt water into a snowcover, II. *Low Temp. Sci., Ser. A*, 1968, **26**, 53-76 (In Japanese with English summary).
4. Wankiewicz, A. A review of water movement in snow. *In* Proceedings of a Meeting on Modelling of Snow Cover Runoff, 1978. Edited by S.C. Colbeck, and M. Ray, US Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 1979. pp. 222-52.
5. Jordan, P. Meltwater movement in a deep snowpack, simulation model. *Water Resour. Res.*, 1983, **19**(4), 979-85.
6. Marsh, P. & Woo, M.K. Wetting front advance freezing of meltwater within a snow cover, 1, Observations in the Canadian Arctic. *Water Resour. Res.*, 1984, **20**(12), 1853-64.
7. Jordan, R. A one-dimensional temperature model for a snow cover: Technical documentation for SNTHERM.89. US Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 1991, Special Report 91-16.
8. Akan, A.O. Simulation of runoff from snow-covered hill slopes. *Water Resour. Res.*, 1984, **20**(6), 707-13.
9. Illangasekare, T.H.; Walter, R.J. (Jr) ; Meirer, M.F. & Pfeffer, W.T. Modelling of meltwater infiltration in sub-freezing snow. *Water Resour. Res.*, 1990, **26**(5), 1001-12.
10. Genuchten, M.T.V. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Sci. Soc. Am. J.*, 1980, **44**, 892-989.
11. Colbeck, S.C. The capillary effects on water percolation in homogeneous snow. *J. Glacoil.*, 1974, **13**(67), 85-97.
12. Denoth, A.; Seidenbusch, W.; Blumthaler, M.; Kirchlechner, P.; Ambach, W. & Colbeck, S.C. Study of water drainage from columns of snow. US Army Cold Regions and Engineering Laboratory, Hanover, New Hampshire, 1979. CRREL Report 79-1.
13. Kattlemann, R. Measurements of snow layer water retention. Proceedings of the Cold Regions Hydrology Symposium, American Water Resources Institute. 1986.
14. Colbeck, S.C. & Anderson, E.A. The permeability of a melting snow cover. *Water Resour. Res.*, 1982, **18**(4), 904-08.
15. Colbeck, S.C.; Akitaya, E.; Armstrong, R.; Gubler, H.; Lafeuille, J.; Lied, K.; McClung, D. & Morris, E. The international classification for seasonal snow on the ground. Issued by the International Commission on Snow and Ice of the International Association of Scientific Hydrology and co-issued by the International Glaciological Society.