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Modeling of Snow Cover Runoff
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CONTROLS ON THE PRODUCTION OF SNOWMELT RUNOFF

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Abstract

The results of two field investigations are summarized; one being in the boreal forest of the Canadian subarctic, and the other in a hardwood forest on the Canadian shield. Runoff during snowmelt is produced in a very different fashion at the two sites. In the subarctic, flow analogous to Hortonian overland flow moves water to surface drainage, whereas in the hardwood forest, flow is dominantly into and through the soil, with insignificant Hortonian overland flow. The major controls on the generation of snowmelt runoff are described for each case, and the relative importance and difficulty of modelling is discussed. These variations in runoff-producing mechanisms between the two type areas have major implications for the development of operational snowmelt runoff models.

This paper is intended to illustrate what may be considered to be an under-researched area in the development of snowmelt runoff models. Flood events produced as a response to snowmelt result from two processes which should be considered separately: snowmelt, and the processes operating on that snowmelt water - the controls on runoff production.

Snowmelt

The best method of predicting snowmelt is to define the energy balance of the snowpack. Under most conditions, the energy balance for an isothermal snowpack may be written as:

$$Q_m = Q_r + [Q_h + Q_e] \quad (1)$$

where Q_m is the heat available for melt, Q_r is the net radiative heat flux, Q_h is the sensible heat flux, and Q_e is the latent heat flux. These two latter terms are generally referred to as the turbulent heat exchanges. Early attempts at defining this (Wilson, 1941; Sverdrup, 1936; U.S.A.C.E., 1955, 1956) were hampered by a lack of knowledge and data, particularly of the turbulent exchange processes. Subsequent developments in micrometeorology (Monin and Obuchov, 1954; Monteith, 1957; Webb, 1965, 1970; Oke, 1970) helped to elucidate the turbulent exchange processes, particularly in the stable conditions which invariably occur over the snowpack. These and other developments have been applied specifically over melting snowpacks, and made possible such works as those of Föhn (1973), Anderson (1968, 1976), de La Casinière (1974), and McKay and Thurtell (1978). The precise definition of the energy balance over a snowpack involves expensive and intensive instrumentation, but a good, working physical model of the energy and mass balance of a horizontal, open snow surface is available.

Variations in the size and relative importance of these three major heatflows occur under various vegetative, topographic, and airmass conditions. Vegetation is an important control. Both the radiative balance and the turbulent exchanges are much modified by the presence of a tree canopy. Petzold and Wilson

(1974) show increased values of net radiation under forest canopies, and Hendrie and Price (1978b) have measured radiative fluxes of >400 W m^{-2} over melting snow in a leafless deciduous forest in eastern Ontario. Net radiation measured in the open over a snowpack of depth greater than 0.15 m, at almost the same time and latitude, and under similar airmass conditions, has maximum values less than 190 W m^{-2} (personal communication, Dr. B. Goodison, 1978). The Ontario data also demonstrate that the turbulent exchanges are highly suppressed in a dense (although leafless) forest - a conclusion taken for granted historically.

Topography causes variations in the receipt of shortwave radiation. This effect is common observation, and the variations in radiative income may be computed from values of radiation measured on the horizontal, using the method of Garnier and Ohmura (1968). Price (1976) applied this method in an attempt to define the energy balance of vegetatively and topographically dissimilar plots in the subarctic. Variations in the shortwave receipt were up to a maximum of 15% for hourly values; this difference being between two plots of high inclination and opposed aspect. Daily totals were not significantly different. In dissected areas the effects of topography on radiative receipt may, however, be of some importance, as noted by Hendrick and Filgate (1971), who modified the work of U.S.A.C.E. (1956) in order to include variability of exposure and vegetative cover.

Generalization about the size and variability of Q_r, Q_h, and Q_e must be made with caution. Price (1976) showed that the notion that Q_r is always dominant in forests is false under certain specific airmass conditions in the boreal forest of Northern Québec. Under these conditions the turbulent exchanges dominated, and in fact produced the largest melt of the 1973 thaw. The results of U.S.A.C.E. (1955, 1956) show that melt in the mountains is largely radiative, and yet under Föhn (Chinook) conditions, Q_h may be much larger than Q_r.

Q_r is relatively easy to measure using a standard Funk-type net radiometer (Funk, 1959). Estimations of the turbulent exchanges, Q_h and Q_e, are much more problematic, involving assumptions concerning the nature of turbulence in the lower atmosphere. They can, however, be reasonably assessed using measurements at one level, as attempted by Price and Dunne (1976), and as outlined by Price (1976) and Anderson (1976). There are uncertainties in this type of estimation of $[Q_h + Q_e]$ resulting from variations in boundary roughness and profile shape. Similarly, for estimates of Q_r, problems may be caused by penetration of shortwave radiation through the snowpack to the ground surface if the snowdepth is less than about 0.15 - 0.20 m (Kondratyev, 1969). However, under most conditions we can make energy balance estimates of snowmelt rates within reasonable limits (Föhn, 1973; Price 1976).

Similar remarks can be made in reference to the application of the degree-day-index (DDI) model (Collins, 1934) to basin estimates of snowmelt. This simple model is often applied because of the lack of any useful data other than air temperature. The DDI model has been applied with some success in many snowmelt models (e.g. Bergström, 1975; Forsman, 1977).

Thus we have serviceable snowmelt models even in the absence of good data. When good data are available, daily or hourly snowmelt can be estimated with some confidence.

Runoff

In the context of this paper, runoff production involves a consideration of all of the processes acting on snowmelt water below the surface of the snowpack. The runoff processes within the snowpack were investigated by Gerdel (1945), and Sharp (1952). The observations of Colbeck and Davidson (1973), and a development of the basic theory of Colbeck (1971) resulted in a physically realistic model of vertical percolation in a homogeneous snowpack (Colbeck 1972). Given estimates of the hydrological properties of the snow, and given the flux rate (melt rate) at the surface of the snowpack, it is possible to predict the form of the input to the base of the snowpack. Subsequently, the Colbeck model was extended (Colbeck, 1974) to include flow over an impermeable ground surface. This model was tested against field data (Dunne *et al.*, 1976), and was found to be a good predictor of runoff from small (2000 m^2) plots. Thus, for the case of a homogeneous snowpack over an impermeable boundary, a working physically based model is available for the prediction of snowmelt runoff. The conditions of zero infiltration and a homogeneous snowpack are, however, extreme cases and the simplest to model. With a structured snowpack, or with infiltration occurring at the base of the snowpack, the situation becomes much more complex. The mechanisms by which water inputs are translated to streamflow involve a debate far from resolved in hydrology, and remain an open question in snowmelt hydrology.

The lateral motion of water from slopes to stream channels derives from four major processes (Figure 1). The most direct and rapid way for snowmelt water to reach the drainage net is the case where infiltration does not occur, or when infiltration capacity is exceeded by input rates. This results in a sheet-flow of water at the ground surface, termed Hortonian overland flow (Horton, 1933). In the context of snowmelt, this type of flow was described by Price (1976). Should the meltwater infiltrate into the ground, then three mechanisms exist which may generate lateral motion. In the unsaturated zone, lateral flow can occur as a result either of changes of permeability with depth associated with textural changes, or, saturation developing above the wetting front (Whipkey, 1965). This runoff component is called subsurface stormflow, or interflow. In the case of dominantly vertical percolation, as might be anticipated in deep, permeable, homogeneous soils, the water will fill soil storage, or reach the groundwater as recharge. This recharge will increase groundwater table slope, and consequently increase rates of baseflow to streams. This type of process has been modelled by Freeze (1971, 1972). Recharge-induced elevation of the groundwater table will increase the size of saturated areas around streams if the water table rises to intersect the ground surface (Dunne and Black, 1970). Any water input onto these saturated areas becomes direct surface runoff, termed partial area contribution, or saturation overland flow. This overland flow is due to saturation from below, which differentiates it from Hortonian overland flow.

In terms of the generation of snowmelt runoff, Hortonian overland flow and partial area contributions form rapidly responding elements in basin reaction to snowmelt inputs, in spite of the presence of the snowpack. Velocities in the order of $34\,\mathrm{m\,h^{-1}}$ within the saturated layer at the base of a snowpack have been estimated

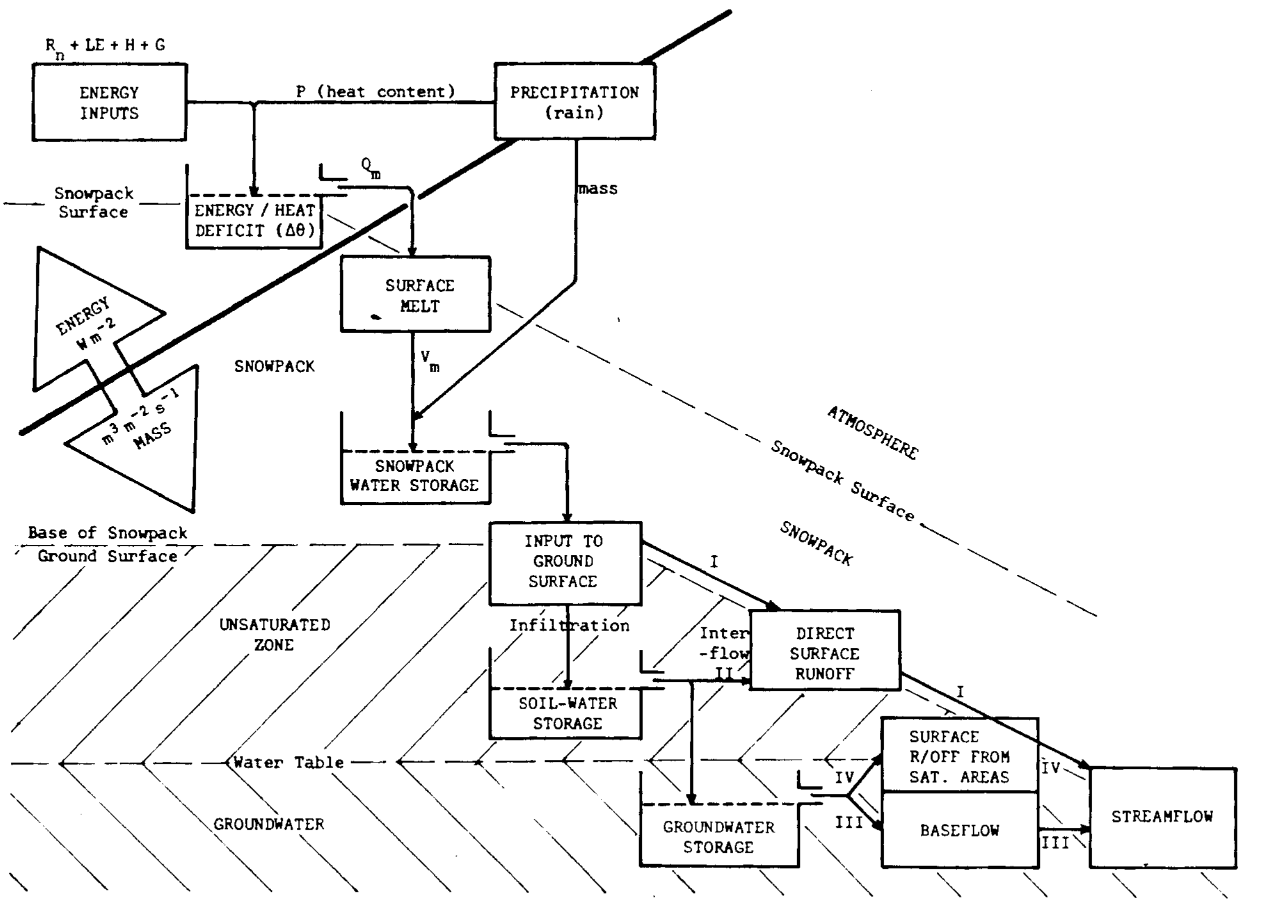

FIGURE 1: Snowmelt water pathways

(Price, 1976). The other runoff pathways involve unsaturated flow, or flow through media of low saturated permeability. In addition, both interflow and baseflow involve major losses to storage necessary to propagate the water pulse, giving long delays and low peaks (Freeze, 1971).

The authors believe that one of the more pressing necessities in snowmelt hydrology is to integrate snowmelt and runoff processes, and to link snowmelt with groundwater fluctuations and to subsequent runoff events, as noted by Colbeck (1977).

Field Studies

The Knob Lake data were collected by A.G. Price and T. Dunne in the thaws of 1972 and 1973 near Knob Lake (Schefferville), Québec, (54° 48'N, 66° 48'W), in the Canadian Subarctic (Figure 2). A full description of the site, and a complete treatment of the methods of analysis and synthesis are given in Price (1976).

The four sites from which most data were collected were in boreal forest of *Picea mariana* (black spruce) and *P. glauca* (white spruce), with an average crown cover of about 15%. An understory of *Alnus* sp. (alder) or *Betula* sp. (birch) is locally developed with much Labrador tea present. The forest floor is covered with a mat of *Cladonia alpestris* (caribou moss) which is about 0.05-0.10 m thick, and is not attached to the mineral soil by roots. The forest soils of the area are developed on thin, patchy glacial till, and are "mini-podsols" (Nicholson, 1973). The surface soil is relatively impermeable even in the unfrozen state. It has a high clay content, and rooting depths are small, generally much less than 0.5 m, resulting in limited organic content and very few root-holes in the upper soil horizons. The soil freezes to some depth in the winter (> 2 m) as "concrete frost" (Post and Driebelbis, 1942; Haupt, 1967). The soil remained frozen just under the surface even when the snowpack had completely melted, and under a 0.15 m snowpack the surface was so solidly frozen that a pickaxe was necessary to pierce it. The hydrological instrumentation on the four forest sites was designed to intercept separately surface runoff and interflow to a depth of 1-2 m. In no case was subsurface runoff observed during snowmelt, and the only place where significant

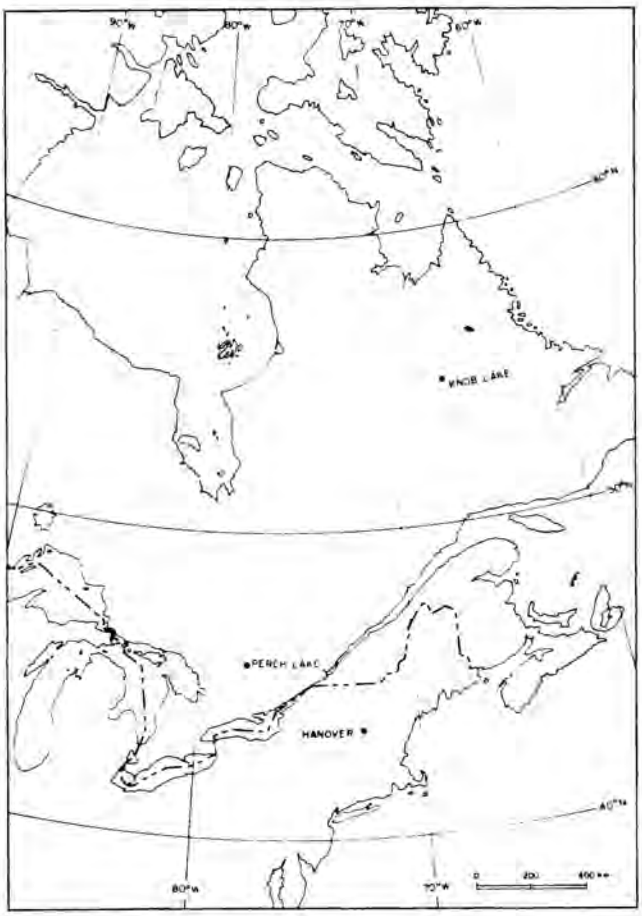

FIGURE 2

Location of measurement sites

subsurface flow was observed was on a steep south-facing slope where small discharges started from the drains several days after the snowpack had completely melted.

During the period of active snowmelt, a marked saturated layer developed at the base of the snowpack. Intercepted runoff was entirely Hortonian overland flow, and the total volume corresponded quite closely with the total amount of snow water equivalent on the plots. These observations confirm the contention that at the scale of 2000 m^2 plots, infiltration below the surface and subsurface runoff were negligible at the Knob Lake sites.

Hourly and daily melt-rates were computed for the plots, the energy balance being compensated for differences in slope inclination and aspect, vegetation cover, and for differing exposures to wind. Insolation was modified to account for slope geometry using the method of Garnier and Ohmura (1968). The mapped shortwave radiation was then used with the regression model of Petzold and Wilson (1974), yielding adjusted net radiation. An Assman psychrometer was used in conjunction with a hygrothermograph to develop a continuous record of atmospheric vapour pressure and temperature, and this record was used with hourly total wind run to make estimates of Q_h and Q_e.

This primitive instrumentation, combined with some theoretical simplifications and plausible assumptions about snow roughness length and exchange coefficients, allowed the computation of daily melt totals for each of the sites, which corresponded well with observed daily runoff volumes.

In the initial period of the 1973 melt at Knob Lake, under an Arctic airmass, the energy balance was dominated by Q_r. In the latter stages of the melt, a large disturbance settled over the area, giving high temperatures, elevated vapour pressures, and gusty winds. During this period $[Q_h + Q_e]$ was much larger than Q_r, and peak daily runoffs were recorded; these large turbulent exchanges existing despite the presence of the forest.

Using the energy balance estimates of surface melt rate, and values of permeability for the snow derived from field observations and the literature, the Colbeck (1974) model was used to predict daily runoff hydrographs for each of the sites. The correspondence between observed and calculated slope hydrographs was good (Dunne *et al.*, 1976), and suggests that the routing model is soundly based.

The variables to which the model is most sensitive are:

(1) Snow depth: The wave of surface flux is distorted and delayed as it passes through the unsaturated zone of the snowpack. The lag and form change increase with snow depth.

(2) Permeability of the unsaturated layer (K_u): This affects the rate and manner of propagation of the wave of water.

(3) Permeability of the saturated layer (K_s): Together with the slope angle and porosity, this defines the rate of travel within the saturated layer, which in turn controls hydrograph peak and responsiveness.

(4) Flux rate at the surface (melt rate): The flux rate affects the velocity of propagation of the wave in the unsaturated zone and the volume of meltwater.

The effect of varying each of these parameters was shown by Dunne *et al.*, (1976). The estimation of K_u and K_s should be improved, but even then the critical parameters are defined by the particular situation. For a shallow snowpack on a gentle slope, the time taken to traverse the saturated layer will be greater than the time taken to get from the surface to the saturated layer. The snowpack effects are therefore small. If the snowpack is deep and slopes steep, as in the Knob Lake case, then the controls exerted by the snowpack will be dominant. In either case the estimates of flux rate (melt rate) are not necessarily the most important parameter. As noted above, the Colbeck model works well only for snowpacks which do not have pronounced horizontal structures. Under these conditions, modelling of flow through the snowpack is much more difficult (Colbeck, 1977). The Knob Lake data define one end of the runoff generation spectrum; the case where infiltration rates are essentially zero.

The second field project was undertaken by A.G. Price and L.K. Hendrie at Perch Lake, Ontario (46° 02'N; 77° 20'W) on the property of Atomic Energy of Canada Ltd. (Figure 2). A description of the site (designated Sub-basin #3) is given by Hendrie (1977). The area of study is within a very extensive forest mainly of *Populus tremuloides* (trembling aspen), *P. grandidentata* (large-toothed aspen), *Betula papyrifera* (white birch) and *Acer rubrum* (red maple). Some *Picea glauca* (white spruce) and *Pinus strobus* (white pine) are present. The primary canopy is dominated by *P. tremuloides*, *P. grandidentata* and *B. papyrifera*, with the secondary canopy consisting mainly of *A. rubrum* with some *Picea glauca* and *Pinus strobus*. There is a brush understorey of *Hamamelis virginiana* (witch-hazel) which is generally distributed through the area. The underlying soil is essentially a deep (> 20 m), permeable sand of high porosity. The litter layer is well-developed and the upper 1 m of the soil profile has much organic material and numerous root-holes. The stratigraphy of the soil is somewhat complicated by the presence of lenses of clayey material distributed non-uniformly in the layer 2-4 m below the ground surface. The infiltration capacity of the soil is not known for the frozen state, but observations suggest values of about $0.10\,m\,h^{-1}$.

The instrumentation at Perch Lake is extensive (Hendrie and Price, 1978b) and estimates of negative heat storage in the snowpack and ground heat flux will be improved next season. The most important portions of the instrumentation for the purposes of this paper are the runoff plots, soil moisture access tubes, and observation wells. Two virtually identical runoff plots of $25\,m^2$ were constructed. One was lined with polythene sheet in order to intercept all melt water reaching the base of the snowpack. The second was left unlined and the measurements from this plot represented just the direct runoff (Hortonian overland flow). Total infiltration was obtained as the difference. Vertical profiles of soil moisture change were determined from measurements taken using the neutron probe in the two access tubes.

Preliminary analysis of the 1978 data allows several generalizations to be made:

(1) On days free from rain-on-snow (or other complicating events), Q_r dominates the energy balance.

(2) The turbulent exchanges are small and positive for most of the melt.

Wind, temperature, and moisture profiles are complex and cryptic, but it is hoped that generalization will be possible when the data analysis is complete. In summary, the energy balance characteristics are simple overall, but complex in detail.

The most striking and informative results of the 1978 Perch Lake investigation are the hydrologic data. The basin, as a system, differs diametrically from the Knob Lake situation, and constitutes the other extreme of the snowmelt runoff spectrum. Originally it was thought that the runoff processes at snowmelt in the Perch Lake basin involved Hortonian overland flow as the major component. The 1977 and 1978 data demonstrate unequivocally that this is false.

The most important controls on snowmelt runoff in this environment are the hydrologic properties of the soil. As noted above, the soil of the Perch Lake basin is a very deep, permeable, porous sand, with occasional included clay lenses. Porosity values of 0.3 to 0.4 are typical of well-sorted sands (Todd, 1959), and the mean value of permeability for the Perch Lake sand is $1.3 - 1.6 \times 10^{-2}\,mm\,s^{-1}$ (Cherry *et al.*, 1975). Flooding of an unvegetated soil surface, of area $3.34\,m^2$, at the experimental site repeatedly gave infiltration rates in excess of $0.1\,m\,h^{-1}$. No determinations of infiltration capacity of the frozen soil were made, but there is strong evidence that it is always greater than water input rates. The continuous litter layer in the forest, and the lack of wash features or channels testify to the absence of overland flow on the basin slopes. Frozen soil conditions might be thought to reduce infiltration capacities substantially, but this has been shown to be false for forest soils. In fact, infiltration capacities under "honeycomb" frost conditions (Post and Dreibelbis, 1942) may be greater than those for unfrozen soil. During snowmelt, input rates to the forest floor are quite low. Under summer conditions, a 50-year storm of 10 minute duration might give a rate of $0.16\,m\,h^{-1}$ ($160\,mm\,h^{-1}$) (personal communication, C.Ll. Morgan, 1978). The peak input rate of the soil surface during the 1978 thaw resulted from a rain-on-snow event, and was $6\,mm\,h^{-1}$). Peak input from snowmelt was $4\,mm\,h^{-1}$ on a day with 20 mm of total melt, meaning that in the spring thaw, rates of input were an order of magnitude less than those possible in summer. The peak rate of snowmelt at Perch Lake was about $20\,mm\,d^{-1}$, and for Knob Lake $60\,mm\,d^{-1}$; a result of the "damping" effect of the tree canopy of the turbulent exchanges.

The rain-on-snow event, mentioned above, occurred early in the thaw. At this time soil temperatures at 0.03 m and 0.08 m showed that the soil was still frozen. Thus, this event gave the largest input rate of the thaw (40 mm in 30 hours) to a frozen soil surface; precisely the conditions under which substantial Hortonian overland flow might be expected. This did not occur. The lined runoff plot yielded 39.1 mm of runoff, and the unlined plot 1.4 mm (3.6% of the input to the ground surface). Although this overland flow component is small, it is almost certainly an overestimate as a result of the nature of the plot construction.

The rain-on-snow event showed that the effect of the snowpack in changing the form of large surface water inputs was minor. For the first sub-event (A, Figure 3), the lag time between input and output was approximately 3 hours, the peak flux rate was reduced by 30% and the volume was reduced by 24%, representing a change in water storage in the snowpack of only 1.2% by volume. Although this change in snowpack storage appears small relative to accepted values of field capacity (4-8% by volume; Gerdel, 1945), in a deep snowpack it represents a large amount of water. For sub-event B, a much reduced lag time, almost identical peak flux rates, and no volume loss were observed. Thus the Perch Lake snowpack was close to field capacity at the end of event A. This decrease in the modifying effects of the snowpack on inputs was probably a result of the filling of field capacity and an increase in crystal size (Colbeck, 1973), referred to as "ripening". Thus, in contrast to the Knob Lake situation, it was possible to ignore the effects of the Perch Lake snowpack upon subsequent inputs.

For the whole melt period the production of runoff from the unlined plot was less than 2.5% of the total snowmelt, so that although Hortonian overland flow exists, it is quantitatively insignificant in terms of total basin runoff. The movement of water within the soil was also investigated. During the 1978 snowmelt, changes in soil water content were derived from measurements made using a Nuclear-Chicago neutron probe at 11 depths in each of two 2 m deep access tubes. The data (Figure 3) were plotted as increases from pre-thaw baseline values in order to avoid the contribution to the count rate due to organic matter. Over the whole period a change in soil water storage of 89.5 mm was determined, and during the same period the total observed input to the soil surface was 91.2 mm. The near equality of these amounts reconfirms the volumetric dominance of infiltration into the soil at the base of the snowpack over Hortonian overland flow at the Perch Lake site. During the period immediately following the rain events, there was an apparent discrepancy between input to the soil surface and changes in the water storage in the soil column. This resulted from the fact that this large input was still close to the soil surface, and accurate measurement of water content in the upper few centimetres of the soil profile is not possible using this technique.

After infiltration, the water percolated down the soil profile at a rate of about 0.2 m d^{-1} (Figure 3). Since this pulse of water is uniformly generated over the basin, it will reach groundwater as recharge first where the water table is nearest the surface. Because these changes are superimposed on a "winter" watertable with low slopes, it means that the first recharge will result in "mounds" of groundwater around lakes and under valleys. This will produce groundwater slopes and flows away from lakes and valleys and towards the areas under ridges; the reverse of "normal" flow. This is of primary importance in terms of runoff production.

It has already been noted that the more rapidly responding types of runoff result from Hortonian and partial area flow. Data suggest that Hortonian flow is insignificant at Perch Lake, so that any rapid response of streamflow to basin inputs must be from partial area flow or direct channel inputs. In order to create saturated areas, the water table must rise to the surface, requiring the storage of large colum s of water. The water table in these groundwater "mound" areas can rise to intersect the ground surface without the necessity of elevating the water table over the whole basin, causing a rapid increase in saturated areas in response to water inputs. This conceptual model of groundwater fluctuations (Hendrie and Price, 1978a) conforms with the observed reversals of groundwater slope at Perch Lake (i.e. away from lakes, streams, and valleys to areas under ridges) early in the melt season.

Groundwater elevation in one well is shown in Figure 3, and can be seen to be time-linked with soil surface water inputs, but is visibly much lagged and damped. The apparent responsiveness of the groundwater table to soil water inputs, suggested by Figure 3, is a function of the shallowness of the watertable at this well location. These interrelated changes in soil water storage, water table elevation, and watertable slope are also reflected in the basin streamflow (Figure 3). The basin hydrograph has two types of variation. The first, a large-scale response to basin

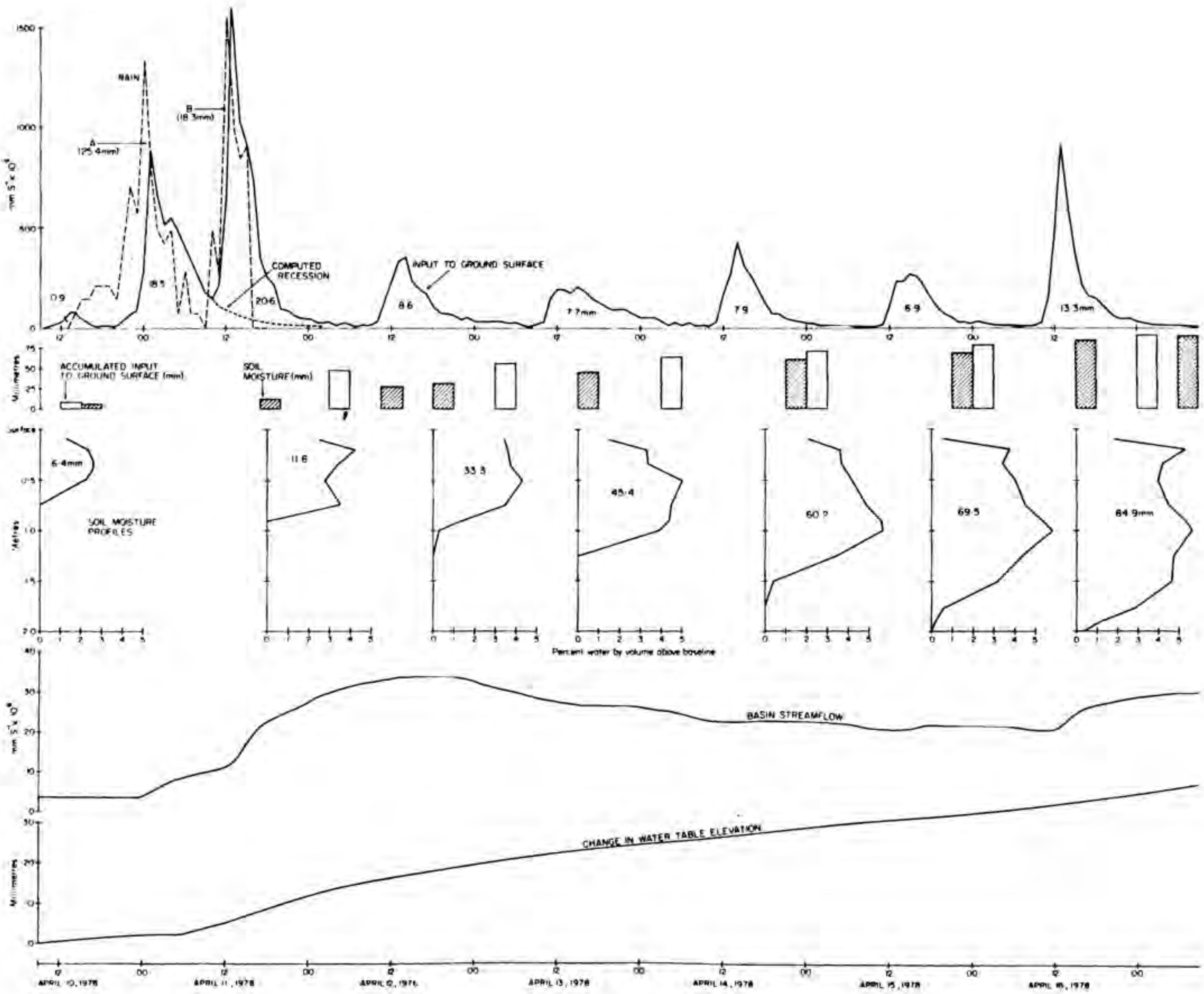

FIGURE 3: Water relationships. Sub-basin #3, Perch Lake, Ontario.

inputs, has a lag-time of 1-2 days, as can be seen from streamflow response to rain events A and B. Superimposed upon this major, slow response is a smaller rise which is lagged by only a few hours from the soil surface input. This small, rapid response can be seen for all days, but gets larger as the melt progresses. If the conceptual model is correct, then the long-term (2 day) response is generated by slow, damped flow through the soil, either as baseflow or interflow. The smaller, rapidly responding element appears to be due to whatever Hortonian flow exists, direct channel inputs, and saturation overland flow from expanding partial areas. The notion of a small but rapidly responding partial area element is strengthened by the observation that the basin streamflow hydrograph shows an increase in the relative size of this "rapid" element later in the melt; this being consonant with the increasing size of saturated areas near surface drainage. This increase is predicted by the conceptual model, and was also observed during snowcourse data collection.

The runoff production processes in the Perch Lake forest are volumetrically dominated by the pathway through the soil, with a smaller, more rapidly responding element generated by saturation overland flow and direct channel inputs. The data show that of the total input to the soil surface, only 10% appeared at the basin outlet, with peak rates of streamflow being about 2.2% of peak input rates. The corollary is that 90% of input remains in storage within the basin, either to re-appear as very lagged streamflow, or to be transpired out of the basin by the forest.

In order to see how well Hortonian overland flow might predict basin streamflow, it was assumed that it produced this 10% outflow. It was also assumed that any infiltrated water remained in storage, or was too retarded to influence streamflow, and that snowpack effects could be ignored. Estimates of porosity and permeability of the hypothetical saturated layer, and mean basin slope inclination and length were made. The Colbeck (1974) model was then used to produce a streamflow prediction from the observed inputs to the ground surface. The hypothetical runoff, and parameters used, are shown in Figure 4 with the observed basin streamflow. A comparison of the two demonstrates that the hypothetical streamflow with Hortonian overland flow as the only runoff-generating mechanism is much too responsive to basin input, and bears little similarity to actual streamflow. Although the assumption of a runoff coefficient of 10% would be volumetrically correct, it is obvious that the effects of runoff-producing mechanisms on the time-distribution of streamflow are primary and overwhelming.

Conclusion

It is recognized that there are gaps in the detailed knowledge of processes of energy exchange over snow which should be filled by specific micro-scale works such as Oke (1970) or Hendrie and Price (1978b). Nevertheless, currently available models (Anderson, 1976) for estimation of snowmelt at a point are satisfactory. At the basin scale, temperature index models have shortcomings under specific conditions (Peck and Anderson, 1977), but generally give adequate results.

In terms of runoff prediction, models exist (Colbeck 1974) which work well in specific circumstances (Dunne *et al.*, 1976), and might be extended to include variations such as inhomogeneous snow (Colbeck, 1977). In any case, in the mid-latitudes, the effects of the snowpack are minor. By far the most important control on the size and form of basin outflows is exercised by interactions at the soil surface. This is particularly true in forests, where soils are porous, usually deep, and permeable even when frozen.

The examples presented here define two extreme possibilities in the generation of snowmelt runoff; no infiltration and total infiltration. Intermediate conditions exist, such as the case investigated by Dunne and Black (1971), where about half of the meltwater infiltrated into the ground.

Although the importance of soil-meltwater interactions has been recognized by snowmelt flood forecasters for some time, the nature of these interactions has remained a major problem. This paper serves quantitatively to illustrate known variabilities. The authors hope the outcome of the Perch Lake project will be a working model for prediction of snowmelt runoff under conditions where infiltration and partial area flow dominate.

Acknowledgements

The fieldwork for both studies was

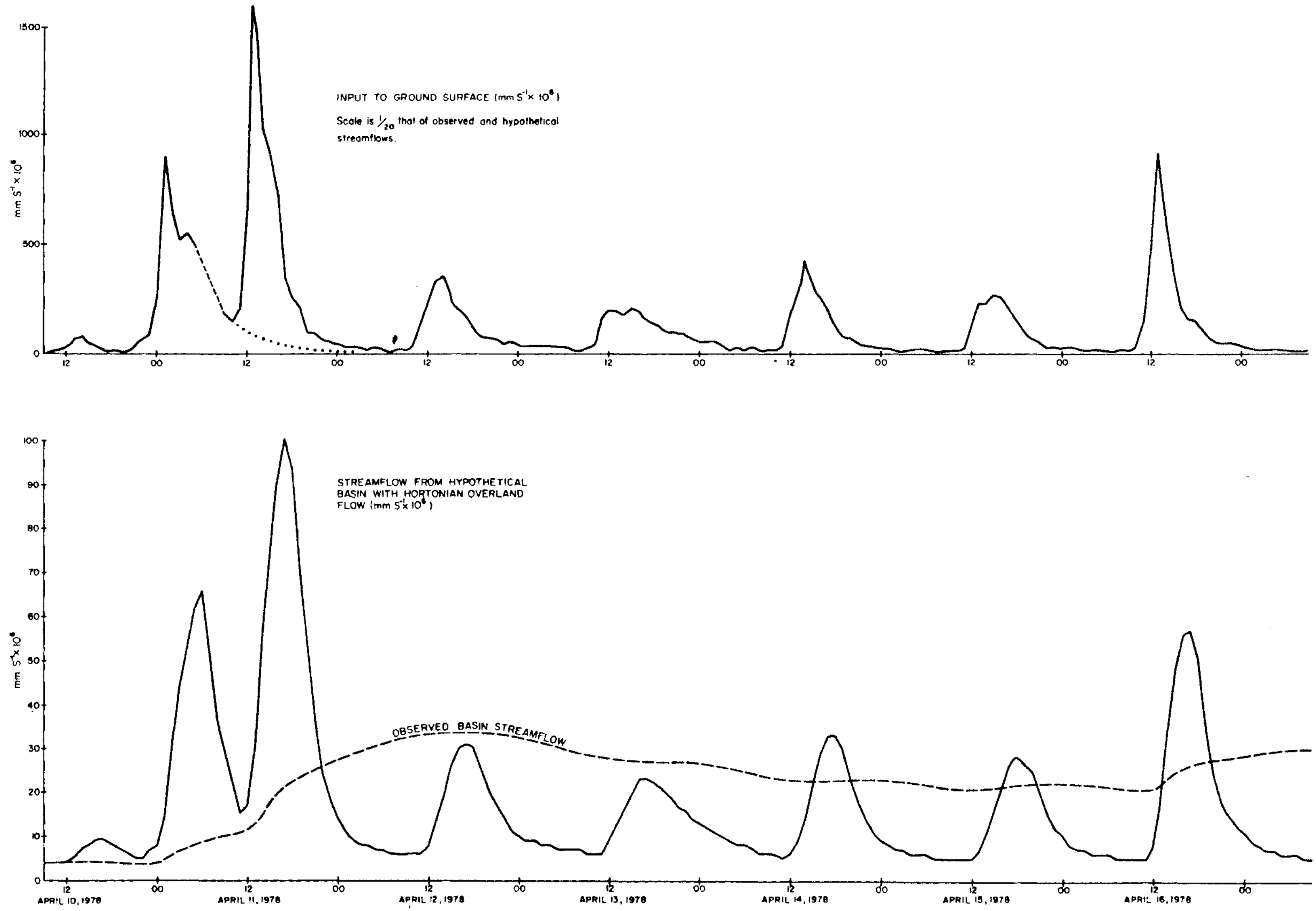

FIGURE 4: Actual and hypothetical basin responses to observed inputs. Sub-basin #3, Perch Lake, Ontario.

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