SNOW HYDROLOGY OF THE HIGH ARCTIC

BY

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INTRODUCTION

The High Arctic of Canada lies well within the Arctic circle (Fig. 1). For about nine months each year, snow and ice constitute an integral part of the landscape, and snow plays a dominant role in the climatology, hydrology and the ecology of the region. Within this vast area, however, there are only four permanent weather stations which routinely record snowfall and carry out snow survey at half-monthly intervals. The extreme paucity and inaccuracy of precipitation data probably forced Lvovitch (1973) to ignore the Canadian Arctic archipelago in his global survey of water balance.

Scientific studies of snow in the High Arctic dated back to the last century. Detailed studies on various aspects of the Arctic snow cover were carried out mainly since the 1950's. Most of these works do not treat the subject comprehensively. This, together with a lack of adequate data base, has not dispelled some of the improper notions regarding the nival processes in a polar environment conceived through cursory observations of early travellers. In the 1970's, a growing interest in the Arctic accompanied mineral exploration and the transport of oil and gas, and a modest body of literature on Arctic snow and ice becomes available.

This paper will review aspects of snow hydrology of the High Arctic, mainly drawing upon works published after 1960, and from field experience of the author and his students in the past decade. Through a survey of the existing knowledge, some insight into future progress may be gained.

STUDY AREA

The High Arctic is composed of many islands. Distances to the coast are small and consequently, few drainage basins exceed 2,000 km^2 in area. Topographically, there is a divergent range of landforms, with plains dominating the western zone, dissected plateaus occupying the central area, and lofty, rugged mountains along the northern and eastern flanks. The entire region was glaciated in the recent geological past, and vestiges of the former ice caps remain in parts of Axel Heiberg, Ellesmere, Devon and other smaller islands.

The climate of the region is strongly influenced by seasonal extremes in the radiation regime. Winter coldness prolongs the duration of the snow and ice cover, and the high albedo of snow greatly limits the net radiation received. Intense and persistent coldness leads to permafrost formation. Continuous permafrost occurs at a depth of about 0.5 m, and total permafrost thickness reaches 400 m (Brown, 1970). These atmospheric and ground thermal conditions are extremely unfavourable for plant growth, so that the High Arctic is essentially a barren region with tundra vegetation occurring only in scattered localities. The lack of vegetation cover in a windswept environment significantly affects the distribution of snow.

SNOW ACCUMULATION

Throughout winter, snowfall accumulation is seldom interrupted by melt events. Except in sheltered locations, however, Longley (1960) found that snow depth at any point does not reflect total snowfall accumulation. Instead, after an initial increase in early winter, depth remains at a certain level which is a function of the local topography and of the direction of the last strong eroding wind. This observation can be related to Tabler's (1975) concept of an equilibrium profile which assumes that any given topography

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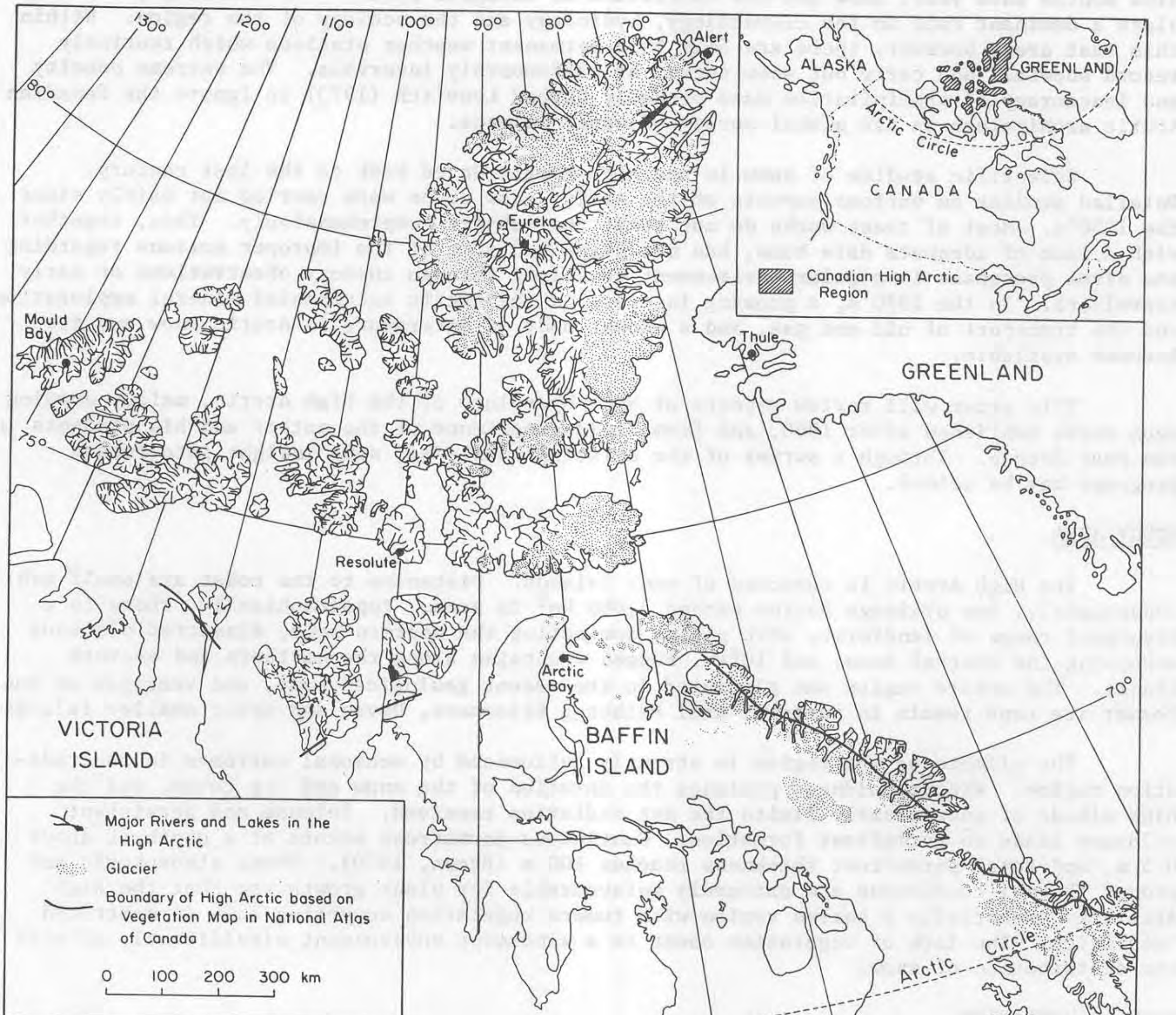

Figure 1. Drainage pattern in the Canadian High Arctic.

has a maximum snow retention capacity. Once this capacity is reached, the snow surface will define the equilibrium profile and further accumulation cannot occur regardless of the amount of snow blowing over the site. This suggests that local terrain controls snow distribution in such areas.

In rugged terrains of the High Arctic, areal variation in the basin snow cover is considerable (Young, 1969). However, it is possible to identify typical topographic units within a drainage basin to obtain meaningful measures of snow depth and snow water equivalent. At McMaster River basin near Resolute, for instance, several major types of terrains can be recognised, including hill-tops, plateaus,lowlying flat areas, gullies, valleys and long slopes of different aspects. Since negligible melting occurs in winter, total snow accumulation in each terrain can be determined by a snow survey conducted immediately before the melt season (Woo and Marsh, 1978). Based on over 500 depth measurements and about 100 densities obtained from each survey, mean values were calculated for different terrain types for six winters (Fig. 2). There is a consistent pattern that depth is the least on hill-tops, followed by flat areas and slopes, with gullies and valleys holding the deepest snow. A year to year variation is also evident, particularly for valleys and gullies. Hence, winters with high precipitation result in a more substantial infilling of the topographical depressions, but for the other types of landform, equilibrium profiles can be easily attained, even during winters of low snowfall.

Given mean depths and mean densities of the snowpack in various terrains, a representative measure of snow storage for a drainage basin can be computed by weighting against the area occupied by different terrain types.

REGIONAL VARIATION OF SNOW PRECIPITATION

Using only four coastal weather station data, attempts have been made to depict the regional pattern of precipitation for the Queen Elizabeth Islands (e.g. Hydrological Atlas of Canada, 1978). Besides insufficient data, the snowfall measurements are open to suspect. Jackson (1960) observed that weather stations underestimate snowfall because of poor catch efficiency of precipitation gauges and because many trace snowfall events are left unrecorded. One proof of the data inaccuracy is the work of Cogley (1975) who found that basin water balance cannot be achieved if weather station precipitation records are used. Another evidence is that the basin-wide snow survey carried out at McMaster River basin produced total winter snow accumulation that was 1.3 to 3 times higher than the total snowfall recorded by Resolute weather station located 5 km away (Table 1). Similarly, the 1980-81 snowfall of Eureka and Mould Bay weather stations was much lower than the snow accumulated in drainage basins nearby. The reliability of the snow survey data is verified by water balance studies. Replacing the Resolute station data with the snow survey data, Marsh and Woo (1979) were able to obtain annual water balances for the McMaster River basin.

Table 1

Comparison of Weather Station Total Snowfall
and McMaster University Snow Survey Data

Location	Date of survey	Snow storage, based on survey (in mm)	Total winter snowfall recorded by station (in mm)
Resolute	May 1976	121	64
	May 1977	95	31
	June 1978	175	112
	May 1979	111	77
	May 1980	107	72
	May 1981	93	70
Eureka	May 1981	67	23
Mould Bay	May 1981	149	51

All values are expressed as snow water equivalent.

Result of the snow surveys also demonstrates that snow constitutes a higher percentage of annual precipitation than was previously recognised. For Resolute area, the weather station data suggest that snow accounts for about 60 percent of annual precipitation

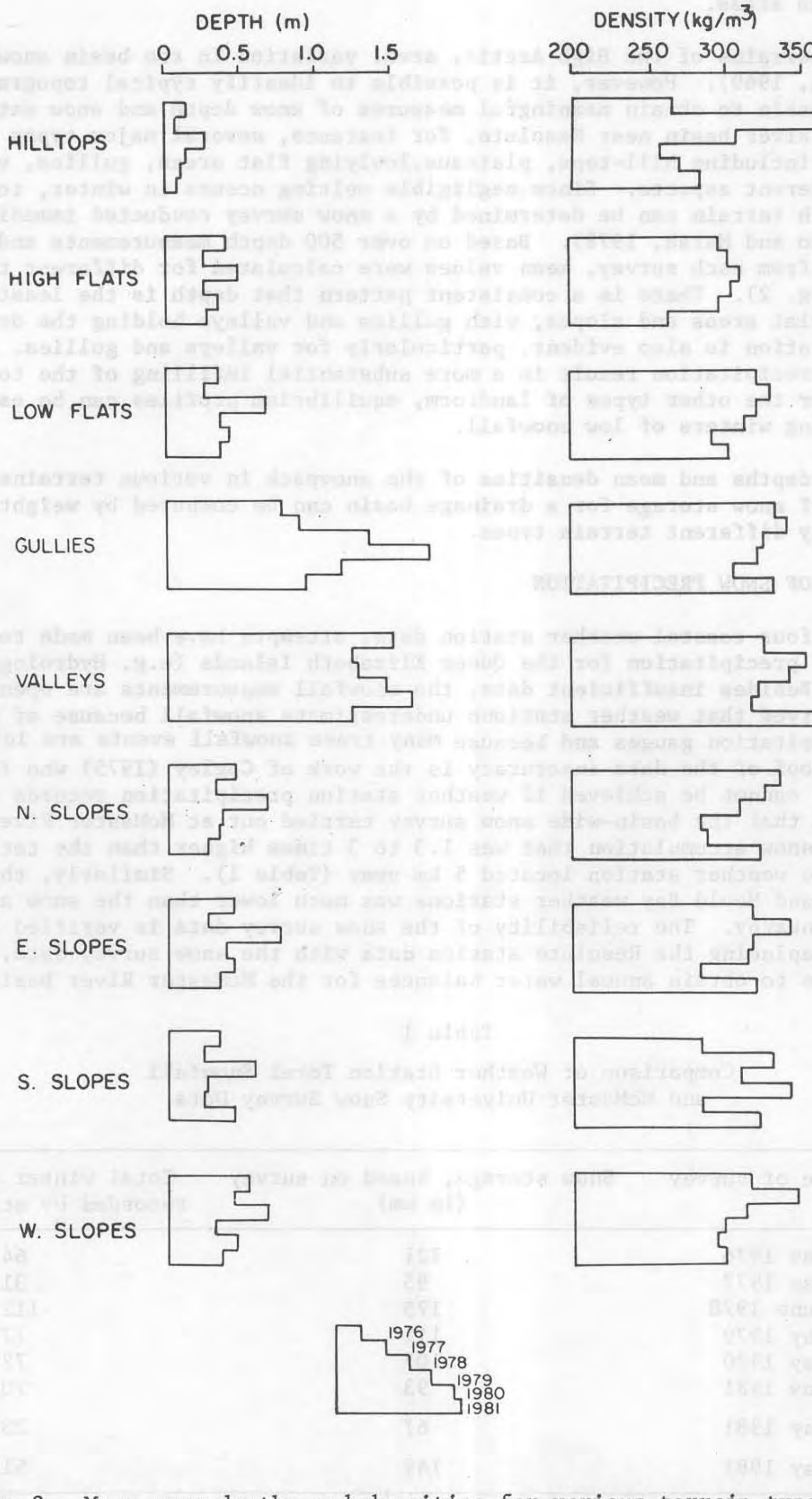

Figure 2. Mean snow depths and densities for various terrain types in McMaster River basin, 1976-81.

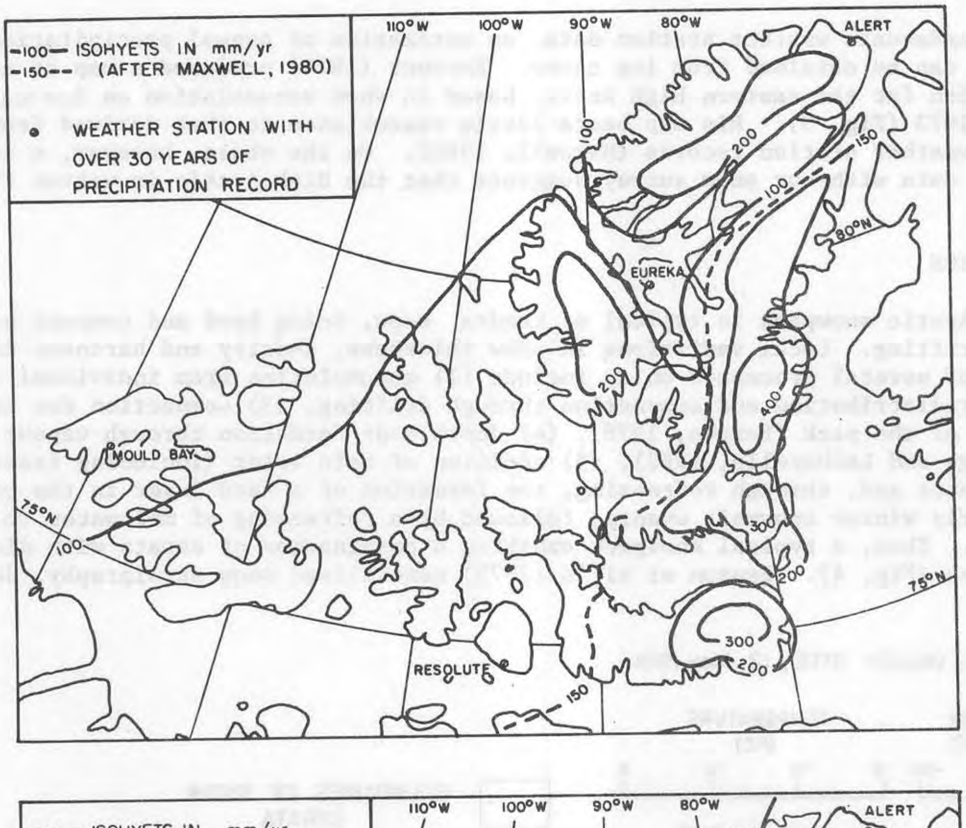

Figure 3. Comparison of precipitation maps for the High Arctic: based on Maxwell (top), and Koerner (bottom).

(Walker and Lake, 1975), but the snow survey measurements revise the snow contribution to about 80 percent.

Despite inadequate weather station data, an estimation of annual precipitation for mountainous areas can be obtained from ice cores. Koerner (1979) produced a map of mean annual precipitation for the eastern High Arctic based on snow accumulation on ice-caps between 1962 and 1973 (Fig. 3). His map bears little resemblance to that derived from an extrapolation of weather station records (Maxwell, 1980). On the whole, however, a combination of Koerner's data with our snow survey suggests that the High Arctic is wetter than we believe.

SNOW CHARACTERISTICS

The High Arctic snowpack is typical of tundra snow, being hard and compact because of considerable drifting. Local variations in snow thickness, density and hardness depend on the intensity of several processes which include (1) accumulation from individual snowfall events, (2) redistribution and compaction through drifting, (3) compaction due to internal pressure of the pack (Bergen, 1978), (4) depth hoar formation through vapour diffusion (Giddings and LaChapelle, 1962), (5) addition of rain water (including freezing rain) in early winter and, through refreezing, the formation of a hard layer in the pack, (6) occasional early winter snowmelt events, followed by a refreezing of meltwater to produce an ice layer. Thus, a typical snowpack exhibits a combination of strata with distinctive properties (Fig. 4). Benson et al.'s (1975) generalised snow straigraphy, derived

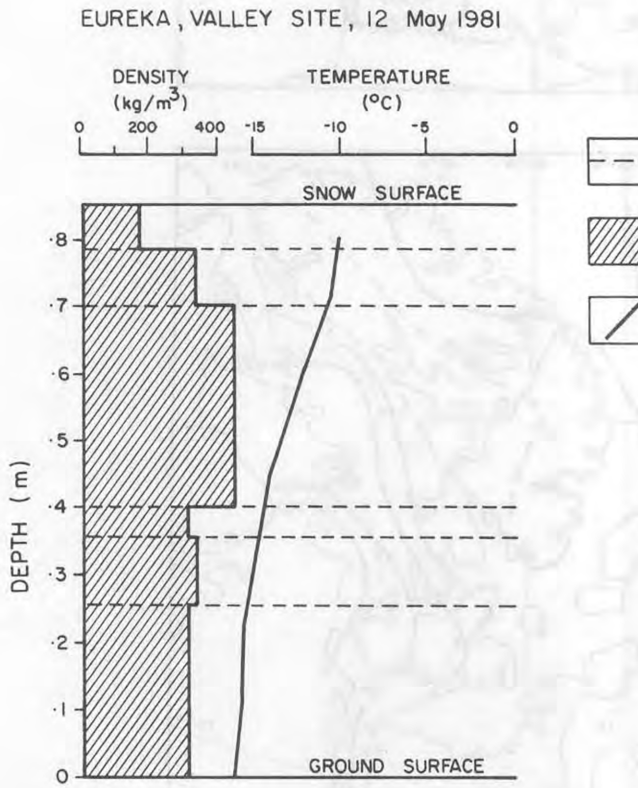

Figure 4. Typical profile showing layering characteristics of High Arctic snowpack (data from P. Marsh).

from observations in Greenland, Antarctica and the north slope of Alaska, can be readily applied to the Canadian Arctic Islands. Four major varieties of snow are recognised, including (1) fresh snow at the surface with variable crystal forms and a density between 150 and 200 kg m^{-3}, (2) hard and fine-grained windslab with a density between 350 and 450 kg m^{-3}, (3) medium-grained snow at a density between 230 and 350 kg m^{-3}, (4) depth hoar consisting of coarse, loosely-bonded crystals yielding an average density between 200 and 300 kg m^{-3}. Depth hoar formation is typical of High Arctic snowpacks. It is particularly noticeable where large vapour gradients in the snow are created by a cold atmosphere and relatively warm substrate.

With low air temperature throughout winter, High Arctic snowpacks are very cold, dropping to below -15°C in late winter. The ground also remains cold, but in early and mid-winter, ground temperatures at 1 m depth can be several degrees warmer than the snow. Such temperature gradient induces an upward vapour flux from the ground to the snow (Santeford, 1978). Although the magnitude of such vapour transfer in the High Arctic is small (Woo,1982) the loss of moisture increases the effective soil porosity, thus allowing some meltwater to infiltrate into the frozen soil during spring.

SNOWMELT PROCESSES

With subfreezing ground temperatures throughout winter and spring, ground melt of High Arctic snowpacks is nonexistent. At the snow surface, the energy available for snowmelt(Q_M) can be considered as

$$Q_M = Q^* + Q_H + Q_L + Q_R$$

where Q^* is net radiation, Q_H is sensible heat flux, Q_L is latent heat flux and Q_R is melt energy produced by rain on snow. The magnitude of various components of the above energy balance varies but some general patterns may be established.

Sublimation (which is a loss of latent heat in terms of melt energy) depends on the dryness of the air relative to the snow surface. Cook (1967) suggested that near Resolute, sublimation is important immediately before the onset of melt events. Cook's comments were strictly observation though later work (Heron and Woo, 1978) found that the quantity of snow thus lost was insignificant. At certain parts of the High Arctic, however, sublimation can be moderate to high. At Truelove Inlet of northern Devon Island, sublimation loss was estimated to be about 20 percent of annual precipitation (Holecek and Vosahlo,1975; Ryden, 1977). In the Expedition Fiord area, Axel Heiberg Island, Ohmura (1972) estimated that half of the heat gained by radiation was spent as latent heat of vapourisation. Observations made at the dry areas near Eureka, Ellesmere Island, show that sublimation produced extensive plough shares on many snowpacks.

Snowmelt is the most intensive between mid-May and mid-June. Radiation melt becomes more important as snow albedo decreases. When turbulent exchange is low, net radiation supplies most of the energy to snowmelt (Woo, 1976). In mountainous areas, high elevation and shaded slopes retain a snow cover for longer durations but in rolling terrains, it was found that slope aspects do not have significant effects on daily total radiation input (Woo and Heron, 1979). This is because all aspects receive sunshine during part of the 24 hour day.

It is common to observe that air temperature increases on clear days. Consequently, both net radiation and sensible heat flux are important sources of melt energy. Towards the end of the melt season, the Arctic snow cover disintegrates into patches because of an uneven initial snow distribution (Woo, Heron and Steer, 1981). Then, advection of sensible heat from the bare ground can accelerate snow ablation.

As a hydrological process, rain-on-snow is not too important because rainfall is usually of low intensity and snowfall rather than rainfall is more frequent during the melt season. When a noticeable rainstorm occurs, however, water penetrates rapidly through the snowpack and this quickens the ripening process.

WATER MOVEMENT THROUGH SNOWPACKS

As a poor heat conductor with a high extinction coefficient for radiation penetration, snow warms up only gradually after May. When surface melt begins, the meltwater percolates through snowpacks with temperatures of -5 to -15°C. The coldness and the layered structure of the Arctic snowpack strongly control the movement of meltwater through the snow.

Meltwater percolation either concentrates along flow fingers or diffuses downward through the pack, the former being a much faster mode of flow (March and Woo, in preparation). Vertical flow is retarded when meltwater encounters an abrupt change in snow structure such as the boundary between adjacent layers. Lateral flow is then facilitated along the layer boundaries. A network of preferred flow routes may be established through a snowpack of moderate depth (1 m, say) several hours after intense surface melt has occurred. Refreezing of meltwater along the network produces ice lenses and ice columns (or glands) some of which measure several mm to several cm in thickness. A special case is when water refreezes at the snow-ground interface to form a basal ice layer (Woo and Heron, 1981).

Continual diffusion of meltwater increases the free-water content of the pack, accompanied by a growth of ice grains through local melt and recrystallization. A release of latent heat due to refreezing of meltwater constitutes the principal mechanism which ripens the snowpack. Note that throughout the ripening period, there is a considerable redistribution of mass within the snowpack but little is lost to surface runoff.

MELTWATER RUNOFF

Runoff often begins after the snowpack is ripe. Due to an uneven basin snow distribution, part of the snow cover may be completely ripen while other areas still retain a partially cold and dry pack. Thus, runoff from the upper slope may flow into a thicker and colder pack on the lower slope and refreeze or be held as liquid water storage. The thinner pack also disappears earlier and slope runoff may move alternately across snow and bare grounds.

Runoff across the snow can take several forms. On the snow surface, unconcentrated flows include the spreading of water over snow (sheet flow) or a downslope movement of water and snow mixture (slush flow). Sometimes, water is concentrated along rills carved in the snow and ice (Woo, Heron and Marsh, 1981). Water movement in the snow includes diffused flow or flow along tunnels and tubes developed inside a pack.

Runoff across bare grounds is controlled by the depth of the thawed zone. Thinly thawed areas are often saturated by abundant meltwater and surface flow is common (Woo, 1976). As the frost table recedes, a thicker thawed layer can accommodate the meltwater supply and subsurface flow then prevails. Locally, rill development is enhanced when surface runoff concentrates along parts of the slopes.

STREAMFLOW

The initiation of streamflow can be delayed for days after the commencement of slope runoff, and this is due to the presence of deep snowpacks choking the valleys at the time of break-up. Unlike the subarctic where river ice is usually present, the amount of ice in the High Arctic streambeds is minimal because of (1) low to negligible winter runoff and (2) limited supply of groundwater to produce icing. Typical break-up events have been described by Pissart (1967), and Woo and Sauriol (1980) related different sequences of events to the pattern of valley snow drift. The initial phase of break-up is a saturation of the valley snowpack, followed by water movement within or on the snow. Several processes then occur, including the formation of channels in the snow, the vertical incision and lateral shifting of channels in the snowpack, the formation and collapse of tunnels carved in deep snow drifts, the ponding and subsequent release of water behind snow dams formed behind drifts (Woo, 1979). The drainage network often develops in segments along the valleys and until the segments are linked, the basin would not have an integrated system of streamflow delivery. This greatly delays streamflow response to basin melt events.

Streamflow measured at most sites along the arctic rivers therefore usually starts with low values, reflecting contributions from a drainage network not yet integrated over the basin (McCann and Cogley, 1972). One notable exception is at the snow-jammed outlets below lakes where initial break-up is associated with severe annual floods caused by a rapid rupturing of the snow jams (Woo, 1980). In most streams, however, high flow occurs only when a local snow jam is breached (Woo and Sauriol, 1981, p.87) or after most segments of the drainage system are united. During the entire period, diurnal cycles of discharge fluctuations are prominent. This reflects the control of snowmelt events, though the lag time between snowmelt and streamflow is not a simple, stationary process (Woo, 1976).

In most basins, high flow lasts for several days to over a week and is followed by a seasonal decrease in discharge (bottom of Figure 5). Such a hydrological regime is called the arctic nival regime by Church (1974) and applies to most streams examined by researchers working in non-glacierized areas (e.g. Ambler, 1974; McCann et al., 1972; Ryden, 1977; Wedel and Way, 1976). It is important to note that all the above works were conducted in coastal basins where the bulk of the basin snow cover disappears by mid-July. There, summer flow is sustained only by suprapermafrost ground water, with high flows produced by occasional rainstorms (e.g. Cogley and McCann, 1976). For basins located at higher

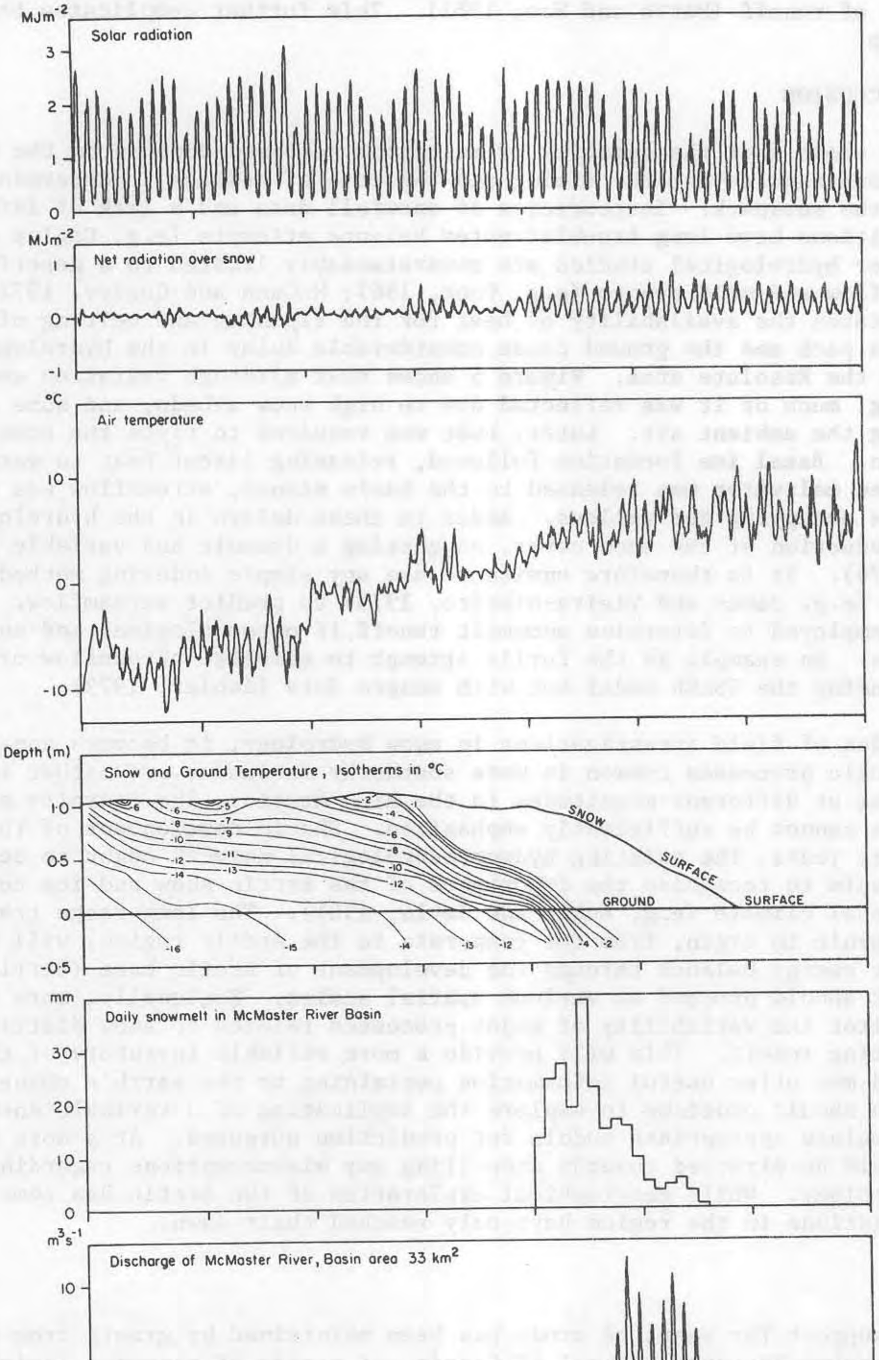

Figure 5. Solar radiation, net radiation over snow, air and snow temperatures, daily snowmelt rate and streamflow of McMaster River. Data collected at or near Resolute, Cornwallis Island, before and during the snowmelt period of 1978.

altitudes, snowmelt is delayed until mid-summer, and the flow regime then resembles the proglacial pattern of runoff (Marsh and Woo, 1981). This further complicates the snowmelt runoff relationship.

DISCUSSION AND CONCLUSION

This paper shows that the quantity of meltwater released depends on the amount and distribution of basin snow cover; the timing and the rate of release is determined by the energy balance of the snowpack. Inaccuracies of snowfall data and a lack of information on basin snow conditions have long troubled water balance attempts (e.g. Cogley and McCann, 1975). Most earlier hydrological studies are understandably limited to a descriptive correlation of streamflow and melt events (e.g. Cook, 1967; McCann and Cogley, 1972). The energy balance dictates the availability of heat for the ripening and melting of the pack, but coldness of the pack and the ground cause considerable delay in the hydrologic responses. An example is from the Resolute area. Figure 5 shows that although radiation energy increases in spring, much of it was reflected due to high snow albedo, and some of it was consumed in heating the ambient air. Later, heat was required to ripen the snow and to sustain sublimation. Basal ice formation followed, releasing latent heat to warm the frozen ground. When meltwater was released to the basin slopes, streamflow was retarded by the thick drifts occupying the valleys. Added to these delays in the hydrologic system is a progressive reduction of the snow cover, suggesting a dynamic and variable source area for runoff(Woo, 1976). It is therefore unwise to use any simple indexing method, such as degree-day factors (e.g. James and Vieira-Ribeiro, 1975) to predict streamflow. Nor should complex models be employed to determine snowmelt runoff if meteorological and snow cover data are inadequate. An example is the futile attempt to simulate streamflow of Falls River, Ellesmere Island, using the SSARR model but with meagre data (Ambler, 1979).

After decades of field investigations in snow hydrology, it becomes apparent that many of the hydrologic processes common in more southerly latitudes are either totally different or operate at different magnitudes in the High Arctic. The scarcity and inaccuracies of snow data cannot be sufficiently emphasized. Should development of the Arctic be inevitable in future years, the existing hydrometeorological network ought to be expanded. More researchers begin to recognise the importance of the Arctic snow and ice cover in controlling the global climate (e.g. Kukla and Gavin, 1981). The long-range transport of aerosols, anthropogenic in orgin, from the temperate to the Arctic region, will have implications on the energy balance through the development of Arctic haze (Barrie et al., 1981). Future work should proceed at various spatial scales. Regionally, more information is required to monitor the variability of major processes related to snow distribution, melt and the resulting runoff. This will provide a more reliable inventory of the Arctic water resources and may offer useful information pertaining to the earth's changing climate. Basin-wide research should continue to explore the implication of a variable snow cover on runoff, and to formulate appropriate models for prediction purposes. At a more fundamental level, efforts should be directed towards dispelling any misconceptions regarding aspects of Arctic snow hydrology. While geographical exploration of the Arctic has come to a close, scientific investigations in the region have only reached their dawn.

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