SNOW ACCUMULATION AND DISTRIBUTION

by

Don M. Gray and Others

INTRODUCTION

Snowcover comprises the net accumulation of snow on the ground as it represents precipitation which has been deposited as snowfall, ice pellets, hoarfrost, glace ice and, in addition includes water from liquid precipitation (rainfall) much of which subsequently has frozen and contaminants stored in the cover. Its structure and dimensions are complex and highly variable with respect to space and time. This variability is a function of many factors: the variability of the "parent" weather, in particular, atmospheric wind, temperature and moisture conditions at the time of precipitation and immediately following deposition; the nature and frequency of the parent storms; the weather conditions between storms during which radiative exchanges may alter the structure, the density and the optical properties and wind action may lead to scour and redeposition as well as modification in snow density and crystalline structure; the process of metamorphism and ablation which can alter the physical characteristics of the snowcover to the extent that it bears little resemblance to the freshly-fallen snow; and topographical, vegetal and physiographical factors. Being the end product of both accumulation and ablation snowcover embraces the complexity of the many factors affecting accumulation and loss.

Prior to embarking on a discussion of snowcover accumulation and distribution it is important to emphasize that the areal variability of snowcover must be considered on three geometric scales. (1) Macroscale or Regional variability, which may include areas from 1 to $1 \times 10^4 \text{km}^2$ with characteristic linear distances of $10^4$ to $10^5 \text{m}$ depending on latitude, elevation, orography, etc. in which the dynamical meteorological effects such as the formation of standing waves, the directional flow of wind streams around a barrier and lake effects are important factors. (2) Mesoscale or Local (within region) variability which may involve characteristic linear distances of $10^2$ to $10^3 \text{m}$ in which redistribution may occur by wind or avalanches along meso relief features and where deposition and accumulation may be related to the terrain variables of elevation, slope and aspect and to vegetal cover variables such as canopy and crop density, tree species or crop type, height of cover and completeness of the vegetal cover.

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Microscale variability in which major differences occur over distances of 10 to $10^2$ m where the accumulation patterns are the result of numerous interactions, primarily the effect of surface roughness conditions as they effect the transport phenomenon.

Inasmuch as snowcover is the residual product of snowfall it is axiomatic that the discussion of its accumulation begin with a brief resume of snowfall formation.

II SNOWFALL (Schemenaur, Maxwell and Berry).

II.1 General Principles of Precipitation Formation

The physics of snowfall is simply a special case of precipitation formation in that the meteorological conditions producing snowfall are the same as those which generate other forms of precipitation. In general, the occurrence of precipitation is determined by the availability of moisture in the atmosphere and the presence of mechanisms capable of converting this moisture into precipitation. Thus, although the formation of snow in the atmosphere depends on many variables, the most important are the presence of supercooled water and an ambient temperature less than 0 °C.

In the northern hemisphere, during winter the oceans particularly those portions of the Pacific and the Atlantic north of latitude 40° N and the Gulf of Mexico, are the main sources of atmospheric moisture at middle and high latitudes. The continental air masses that move over the oceans generally have a much lower temperature and vapor pressure than the underlying water surface. As a result large upward fluxes of heat and moisture to the lower layers of the atmosphere occur. The extent to which this moisture is transferred to higher levels depends both on the nature of the surface energy exchange processes and on the large scale vertical motion of the atmosphere which is influenced by the configuration of the upper atmospheric flow. Although on a hemispheric scale oceans serve as the principal sources of moisture supply other bodies of water can be important particularly in the early parts of winter before they freeze. For example in North America Hudson's Bay and the Great Lakes.

The amount of water vapor present in the atmosphere in a particular region is influenced by both the temperature of the air, which determines the upper limit of its vapor capacity and the accessibility of source areas of moisture. When considering the broad-scale distribution of atmospheric moisture it is common to use the concept of precipitable water, the depth to which water would stand if all water vapor were condensed from a vertical column of uniform cross-section of the earth's atmosphere. Hay (1970) has prepared maps of the mean precipitable water for different months for Canada, the distribution for January is given in Figure 1. The largest values for the month (8 to 12 mm) which occur along the British Columbia coast can be attributed to warm moist air originating from the Pacific. A second maximum of 4 to 7 mm which occurs over the Atlantic Provinces reflects the effects of the Atlantic and, to a lesser extent, the Gulf of Mexico on the atmospheric moisture supply. Throughout much of the remainder of Canada the values fall in the range from 2 to 4 mm and decrease with increasing latitude. This decrease can be attributed to the low temperatures experienced by those regions in January and to the fact they are distant to major moisture sources.

Precipitation normally occurs when the vertical motion in the atmosphere causes cooling of elements of an air mass by adiabatic expansion (an expansion process is termed adiabatic when there is no heat exchange with the surrounding mass). Four types of vertical motion associated with the formation of significant amounts of precipitation are:

1. Horizontal convergence: wind fields are oriented to direct the flow into a particular area thereby causing lift.
2. Orographic lift: when the movement of air is directed
against a ridge, group of hills or mountains and air is forced to rise.

(3) Convective lift: lift produced by differential heating or advection which causes a portion of the atmosphere to become more buoyant than its environment.

(4) Frontal lift: lift produced by the interaction of separate large masses of air having significantly different physical characteristics (density and motion).

Of these four lifting mechanisms the last three are most important to snowfall formation. In North America the effects on snowfall accumulation originating from orographic effects are well recognized in the mountainous regions adjacent to the Pacific Ocean. Likewise, major accumulations of snowfall resulting from convective activity are common in areas south and east of Lake Huron and Lake Ontario. Frontal storms tend to give widespread snowfall to large areas of the interior parts of Canada and the United States. In the absence of instability, the precipitation patterns originating from frontal activity show steady precipitation within 250 km ahead of a warm front and within 40 to 80 km of the cold front. However, conditions may vary markedly in any given storm depending on the moisture supply and the vigor of the frontal wave. Strong convective activity indicative of instability complicates the pattern and leads to the occurrence of pockets of heavy precipitation. Figure 2 illustrates the typical cloud structure through warm and cold fronts. Correspondingly, it is possible to characterize the snowfall associated with the different cloud types (see Table 1).

II.1 Physics of Formation

The preceding section provides a very brief resume of the macroscopic factors affecting the formation of precipitation. It is important however to consider in greater detail the physics of snowfall as the formation process governs not only the depth or amount of snowfall but also the physical properties of the snow crystals such as their shape, mass, density, crystal habit, etc. The snow crystal that arrives at the ground may be a simple product of ice crystal formation or it may be the result of a complex life history during which its original physical character is greatly modified and changed.

The formation of snow in the atmosphere depends on many atmospheric variables, the most important being that the ambient temperature is less than 0 °C and that supercooled water is present. The schematic diagram shown as Figure 3 outlines the mechanisms by which different types of snow develop.

At a temperature of -5 °C ice-forming nuclei present in the atmosphere form

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**Figure 2.** Typical cloud structure associated with warm and cold fronts. (cloud abbreviations given in Table 1.)
Table 1. Snowfall associated with various cloud types

<table>
<thead>
<tr>
<th>Cloud Type</th>
<th>Associated Snowfall</th>
</tr>
</thead>
<tbody>
<tr>
<td>Typical Cirrus and</td>
<td>Precipitation is usually virga although light snow showers are possible from the</td>
</tr>
<tr>
<td>warm derivatives (Ci)</td>
<td>Acc</td>
</tr>
<tr>
<td>frontal cloud</td>
<td>Light continuous or intermittent snow may fall from As; however, when the snow is</td>
</tr>
<tr>
<td>Altocumulus (Ac)</td>
<td>heavy, the As has probably graduated to a nimbostratus-type cloud.</td>
</tr>
<tr>
<td>Castellanus (Acc)</td>
<td>Continuous snow may fall. Virga occurs from both Ns and As.</td>
</tr>
<tr>
<td>Altostratus (As)</td>
<td>Intermittent light powdery snow (fine flakes) possible.</td>
</tr>
<tr>
<td>Nimbostratus (Ns)</td>
<td>Continuous light powdery snow is possible. (This is the frozen precipitation analog of</td>
</tr>
<tr>
<td></td>
<td>warm precipitation drizzle.)</td>
</tr>
<tr>
<td>Stratocumulus (Sc)</td>
<td>Light snow showers possible although these are more likely to occur from the Tcu.</td>
</tr>
<tr>
<td>Stratus (St)</td>
<td>Moderate to heavy snow showers are possible.</td>
</tr>
<tr>
<td>Typical Cumulus</td>
<td></td>
</tr>
<tr>
<td>cold Towering</td>
<td></td>
</tr>
<tr>
<td>fronted cloud</td>
<td></td>
</tr>
<tr>
<td>Cumulus (Tcu)</td>
<td></td>
</tr>
<tr>
<td>Cumulonimbus (Cb)</td>
<td></td>
</tr>
</tbody>
</table>

tiny ice crystals through the process of ice nucleation. The nucleation of the ice phase on these particles or nuclei is called heterogeneous nucleation which is the primary process leading to the formation of ice crystals. Three basic types of heterogeneous nucleation may occur: (a) if an ice nucleus is present in a cloud droplet which is cooled below 0 °C the droplet will freeze at a temperature determined by the nature of the nucleus, in which case the ice nucleus has acted as an "immersion freezing" nucleus; (b) if an ice nucleus touches the outer surface of a droplet and causes it to freeze, it has served as a "contact nucleus"; and (c) if water vapor is deposited directly on a particle and eventually forms an ice layer on the surface the particle has acted as a "deposition nucleus". The nucleation processes are extremely complex.

An ice crystal in a cloud of water droplets grows at the expense of the droplets because the vapor pressure at the ice surface is less than that above the water surface. At temperatures conducive to the formation of snow a cloud

\[
\text{water vapor} + \text{ice nuclei} + \text{cloud droplets at } T < 0 \degree C \\
\text{ (nucleation)} \\
\text{ICE CRYSTALS} \\
\text{(sublimation)} \\
\text{(riming)} \\
\text{SNOW CRYSTALS} \\
\text{(sublimation)} \\
\text{(aggregation)} \\
\text{RIMED CRYSTALS} \\
\text{GRAUFEL} \\
\text{SNOW FLAKES}
\]

Figure 3. Schematic flow diagram of the formation of different types of snow.
may exhibit only slight supersaturation with respect to water and supersaturations of 10 to 20% with respect to ice. The net result is a transfer of water vapor from the droplets to the surface of the ice crystal. The preferential growth of the ice crystals is the basis of the Bergeron mechanism (Bergeron, 1935) of precipitation formation, a theory which describes the development of much of the precipitation in the temperate latitudes. When the crystals attain sufficient mass they fall and during this movement grow primarily by a riming process. If during its fall the ice crystal passes through a layer of the atmosphere in which the temperature is greater than 0°C it may melt completely and fall as rain. The basic habit (shape) of an ice crystal is determined by the temperature at which it grows whereas the rate of growth and secondary crystal features are determined by the degree of supersaturation. Further different relative growth rates of the crystal faces leads to extremes to plate-like or prism-like crystals.

The rate of change in mass of a crystal by diffusion is directly related to the difference in vapor density of the atmosphere and the air at the surface of the crystal. When a crystal reaches a size of a few hundred of μm its mass growth by diffusion becomes relatively less important to accretion or the capture of droplets through collisions. The riming process is important in that riming affects the fall velocity and motion of the crystals, it is through the riming process that graupel and most hailstones originate. Recent experimental evidence has indicated that during freezing of a cloud droplet onto a snow crystal secondary ice particles are ejected leading to a "multiplication" of ice particles in a cloud. The onset of riming occurs at different sizes for crystals of different shape.

A snowflake is formed by the aggregation or the collision of snow crystals followed by their adhesion. Adhesion can result by the interlocking of parts of the crystals, riming, vapor deposition and sintering. The maximum sizes of snowflakes generally occur at temperatures near 0°C. Hobbs (1974) shows that in a typical cloud a 1 mm diameter snowflake can grow to 10 mm in about 20 minutes.

The process governing the formation of a snowflake in the atmosphere as well as affecting the crystal habit also governs the sizes, masses, the densities, the terminal velocities and the concentrations of snow crystals. Individual snow crystals observed at the earth's surface range in dimension from ~50 μm to ~5 mm. Unfortunately, the density of snow crystals of a given type is difficult to specify because of the problems in determining a representative volume. The particle densities for heavily rimed crystals and for graupel, whose volumes can be measured with reasonable accuracy have been found to range from approximately 100 kg/m³ to 700 kg/m³.

Snowflakes can consist of from two to several hundred snow crystals joined together. They may range in size (maximum diameter) from 0.10 mm to several cm. The largest snowflakes occur at temperatures near 0°C with decreasing sizes as the temperature decreases, (Hobbs, 1973). Magnano and Nakamura (1965) showed that the density of a snowflake decreased with size according to the relationship:

$$\rho d^2 = 2 \times 10^{-6}$$

where \(\rho\) is the particle density in mg/m³ and \(d\) is in mm.

A knowledge of snow formation in the atmosphere is essential to winter weather modification activities including the artificial redistribution of snowfall and the initiation and increase of snowfall from a given cloud system.

II.2 Areal Distribution

Fulkes (1935) derived the following approximate relationship for the precipitation rate of an ascending layer of saturated air of unit cross section:

$$i = bw\Delta z$$

where: \(i = \) precipitation rate,
\(b = \) a coefficient whose magnitude depends on the temperature and pressure of the layer,
\(w = \) the vertical velocity, and
\(\Delta z = \) the thickness of the layer.

The variation in the coefficient \(b\) with temperature and pressure height is shown in Figure 4. As shown in the figure under conditions when most snow forms (temperatures below 0°C and heights below 6 km), the rate of precipitation decreases rapidly with decreasing temperature; at temperatures of ~10°C and lower the rate is relatively uninfluenced by height. The vertical velocity is determined primarily by the characteristics of individual weather systems and the extent to which
terrain factors affect the airflow. The duration of snowfall depends on a number of factors including the speed and track of the system and the size and shape of the associated snowfall area. In the case of convective storms, the intensity of snow formation depends on the availability of moisture and the degree of instability of the air. Convection which produces snow often occurs with the movement of cold air over warm bodies, for example, lake effects. The amount by which the moisture and stability of the air is modified depends on factors such as the initial temperature and moisture characteristics of the air, the temperature of the water and the length of overwater trajectory. Instability can also be produced by large-scale motion associated with frontal systems.

II.2.1 Effects of Moisture and Precipitation Mechanisms on Snowfall Distribution

The distribution of the mean annual snowfall amounts over Canada and the United States is illustrated in Figure 5. In the figure, two areas of relatively heavy snowfall can be identified. Certain parts of western British Columbia, the Yukon and Alaska adjacent to the mountain ranges which parallel the Pacific Coast, receive seasonal values exceeding 400 cm. These areas are directly exposed to moisture-laden disturbances moving eastward from the Pacific source region. The vertical motion associated with low-pressure systems is enhanced by the coastal terrain and the snowfall can be very heavy, in some locations exceeding 1000 cm annually. For example, along the southern coast of British Columbia near sea level, the air temperatures are normally above freezing so that a majority of winter precipitation occurs as rain, and the seasonal average snowfall is less than 60 cm. Snowfalls are also relatively light in areas to the lee of the mountains where eastward moving air has a downward component.

Widespread heavy snowfall also occurs in Eastern Canada throughout an area encompassing central Ontario, southern Quebec, much of the Atlantic provinces, Labrador, and the east coast of Baffin Island where the seasonal falls range from 250 cm to in excess of 400 cm (see Figure 5). Moisture is available to these areas in varying amounts from the Pacific, the Atlantic and the Gulf of Mexico. In addition, to the west, the Great Lakes serve as an important source of moisture for local precipitation. For example, average seasonal snowfalls in excess of 250 cm occur to the southeast of Lake Huron. The snowfall amounts decrease rapidly in the southward direction from the eastern Ontario-northern New England area to the southeastern United States. For the most part, this is a result of increasing temperatures, as opposed to decreasing precipitation.

Over the prairie provinces and the northern plains the seasonal snowfall is considerably lower than in the eastern or western regions of the country, averaging between 70 to 140 cm in most areas (see Figure 5). The small amounts of snowfall over this region can be attributed, in part, to the infrequent occurrence of vigorous weather systems. Likewise, the relatively flat terrain does not induce snowfall formation as the Pacific air moving inland tends to subside as a result of the downward slope in topography from the Rocky Mountains to Manitoba-Minnesota.

The western half of the Arctic Islands receives less snow (<80 cm) than most other parts of Canada. Although winter is long in this area, it is remote from major moisture sources, and the extremely low temperatures which prevail over the region reduce the moisture holding capacity of the air to low values.
III FACTORS AFFECTING SNOWCOVER ACCUMULATION AND DISTRIBUTION

III.1 Wind (Kind)

The characteristics of the wind near the earth's surface are of major importance in determining the amount of snow movement and in determining the scour and depositional patterns. In addition, when snow crystals are moved by wind, their physical shape and properties may be changed markedly.

At heights greater than about 1 km above the earth's surface the motion of the atmosphere is governed by the pressure distribution of large-scale weather systems and the wind characteristics are independent of terrain conditions at the surface. Throughout the boundary layer the wind speed increases from zero at the earth's surface to the geostrophic wind speed at its outer limit. The thickness and distribution of the velocity profile depend on the nature of the earth's surface. Numerous studies have been conducted on the influence of different surface conditions, particularly vegetative cover, on the wind velocity distribution with height (see Figure 6). For detailed information on these effects the reader is referred to the works of Geiger (1966) and Reifsnyder (1955). Over relatively rough terrain the boundary layer is thicker and the wind speed increases relatively slowly with height; over flat, open terrain the opposite is true. The boundary layer thickness and velocity profile are also influenced by the
vertical wind profiles in different forest stands (after Reifsnyder, 1955).

The presence of velocity gradients within the boundary layer implies the existence of shear stresses in the wind flow. The shear stress is a maximum at the earth's surface and decreases with height becoming zero in the geostrophic wind above the boundary layer. It is the shear stress exerted by the wind on the surface which causes the movement of loose snow. Virtually all natural surfaces act as rough surfaces with respect to wind. Under these conditions, and when the wind speed is high enough to induce drifting or movement of snow the flow near the surface is dominated by the shear forces and the effects of thermal stratification are negligible. Under such conditions field observations have shown the following relationship applies:

\[ \frac{U}{U^*} = 2.5 \ln \left( \frac{z}{k} \right) + 5.5 - C(\lambda) \]  

(3)

where: \( U \) = the mean wind speed at a height, \( z \),
\( U^* \) = the "friction" or "shear" velocity, equal to \( \sqrt{\frac{\tau}{\rho}} \) in which \( \tau \) is the shear stress at the surface and \( \rho \) is the density of air,
\( k \) = a roughness parameter; and
\( C(\lambda) \) = a constant whose magnitude depends on the non-dimensional spacing of the roughness elements, \( \lambda \).

The maximum height to which the logarithmic velocity profile equation (equation 3) applies is limited by the upstream fetch over which the surface roughness is reasonably uniform. That is, Tani and Meketa (1971) suggest that the equation is valid for \( z \leq \text{fetch}/20 \) or \( z = 50 \text{ m} \), whichever is least. The lower limit of applicability of the equation is for \( z \leq 2k \). That is the equation will not describe the profile near or below the tops of roughness elements where the flow pattern is very complex and often three dimensional, for example, where bushes, trees, buildings and other obstacles affect the wind pattern.

The limitless variety and combinations of surface features that can be encountered in nature make it impossible to define and to discuss all the interactions between air flow patterns and snowdrift patterns. The snow accumulation patterns are a complex function of deceleration and acceleration of the air stream, its velocity profile and the formation of separation bubbles and vortices (Richter, 1945).

III.1.1 Transport of Snow by Wind

Individual particles of snow lying on the surface are initially caused to move by the drag force exerted on them by the moving air. The drag force per unit area is the shear stress, \( \tau \), existing between the moving air and the snowcover, which, in turn, is a function of such factors as surface roughness, the mass density of air and the wind velocity. Before movement can occur it is necessary that the shear stress attain some critical value to overcome the particle weight and inter-particle cohesive forces; at snow temperatures less than about \(-2{\circ}\text{C}\) the cohesive forces in newly-fallen snow may be considered negligible. Normally, in snow drifting studies, it is customary to utilize the shear velocity, \( U^* = \sqrt{\frac{\tau}{\rho}} \), instead of the shear stress. The threshold shear velocity \( (U^*)_h \) required to disturb the surface and transport particles is highly variable depending on the size, shape and weight of the snow crystals and the cohesive forces, the latter being dependent on the wetness of the snow. These aspects stress the importance of the
formation process with respect to snow drifting. Numerous studies (Richter, 1945; Kotlyakov, 1961; Oura et al., 1967; and Isyumov, 1971) have been directed to evaluating $U^*$ for different snow conditions. The values of $U^*$ found in these studies vary from 0.07 m/s for a light, dry snow to 1.0 m/s for wind-hardened snow of similar density (350-400 kg/m³) but differing hardness (98 kPa compared to 245 kPa). In fact, the data reported by Kotlyakov (1961) for the Antarctica show a linear relationship between snow hardness and $U^*$. This relationship is complicated by the fact that in the absence of new snowfall, compaction by wind and energy exchange processes lead to a hardening of the snow surface. It is interesting to note that Oura et al. (1967) report an increase in $U^*$ of newly fallen snow from 0.22 m/s to 0.4 m/s after only several hours of aging.

The shear stress, $T_s$, and the shear velocity $U_s$ are related to the wind velocity, $U$. Owen (1964) and Kind (1976) suggest that the following relationship can be used to describe the velocity profile over and within the saltation layer

$$\frac{U}{U_s} = 2.5 \ln \left[ \frac{z}{U_s^2/2g} \right] + 9.7 \quad (4)$$

Obviously, equation 4 is the same as equation 3 provided $U_s^2/2g$ is interpreted as being proportional to the effective roughness height, $k$. Mellor (1965) provides an extremely useful summary of the roughness of snow.

### III.1.1.1 Modes of transport. The three modes of snow transport recognized in the movement of snow are: ground creep - the physical movement or rolling of particles along the surface saltation - the bouncing of particles along the surface travelling in a curved trajectory under the influence of wind and gravity forces and turbulent diffusion - in which particles are held in suspension in the air stream without necessarily contacting the ground. Generally, it is accepted that most of the snow is transported by saltation or turbulent diffusion. In effect, these two modes of transport have led to the development of two snowdrift theories, the dynamical and diffusion theories based on the works of Bagnold (1941) and Schmidt (1925). As detailed by Radok (1977) the basic equations of the two theories are of somewhat marginal relevance from a practical point of view. In essence, the most telling difference is in terms of the dominant processes and vertical scales, the dynamic theory views snow drifting as a near surface phenomenon due to small eddies in the lowest 10 cm producing mainly saltation, the diffusion theory (relating to conditions on polar ice sheets) attaches the main importance to the larger eddies in the free air stream extending to tens or even hundreds of metres above the surface. Radok (1977) in his evaluation of the two theories stresses that the real power of the diffusion theory comes from its detailed predictions of drift concentrations and velocity profiles and a greater understanding of the snow drift process.

In practice snow particles can only rise to great heights above the earth's surface when the vertical velocity components are approximately equal to or greater than the terminal falling velocity, $w$, of the particles. Bagnold (1973) suggests that the turbulent velocities become roughly equal to the terminal falling velocity of particles only when the shear velocity $U^*$ is greater than about 5 $U^*$. Generally, the dominant particle size (nominal diameter) of a wind blown snow may be of the order of magnitude of 0.5 mm for which values of $U^*$ may range from 0.1 m/s to 0.2 m/s. Winds capable of producing a shear velocity of 0.74 m/s (corresponding to approximately 14 m/s at a height of 1 m) are infrequent in nature. In recent years considerable study has been devoted to the mass transport of snow. Owen (1964) has shown that for saltating uniform, spherical particles, assuming no phase change and two dimensional flow the mass transport rate or the ability of the wind to transport snow is approximately proportional to $U^*$. However, it is recognized that during the scouring or erosional process layers of very fine particles (snow dust) may be exposed which, because of their low fall velocities, may rise to great heights in the atmosphere and become suspended in the turbulent air flow. Kind, using field data reported by Oura (1967) and Kobayshi (1973) plotted a curve showing the distribution of the horizontal mass flux of blowing snow with height (see Figure 7). It is evident from the figure that approximately 90% of the total flux occurs within about 2 cm of the surface (saltation).

### III.1.2 Condensation/sublimation Losses.

Obviously, were it possible to define mathematically the distribution of
snow particle concentration the mass flux would be calculable by integrating the function with height. Unlike solid particles, when a snow crystal is transported, its mass changes because of vapor transport with the surrounding air and other particles. Schmidt (1972) presented a model of the major processes contributing to the evaporation or sublimation of a single particle. Some important findings from this work included: the sublimation rate appeared to double for each 10 degree temperature rise in the range from -20°C to 0°C. It more than doubled when the particle diameter doubled and the percentage mass loss in unit time increased markedly with decreasing particle size. In essence, a snow particle evaporates because of vapor pressure gradients existing with the surrounding air and other particles. Schmidt assumed that all the sublimated vapor is transferred vertically by turbulent exchange hence it would be hypothesized that a substantial amount of water to a snowcover may be lost in the transport process. Equations for estimating the condensation/sublimation loss of a mass of blowing snow have been proposed by Dyunin (1961) and Tabler (1971). In both cases use is made of the concept of a "transport distance", Ru, the distance over which the average sized drift particle is completely evaporated. Obviously, the transport distance frequently differs from the contributing distance and thus the amount of blown snow completely collected by some barrier or obstacle must be adjusted by the amount which has evaporated. In a later paper Tabler (1975) suggests that the ratio between the percentage of relocated precipitation lost to evaporation as a function the ratio of the fetch to the transport distance is an exponential function - derived from a consideration between evaporation and particle size distribution. For conditions encountered in Wyoming where environmental conditions were constant over the fetch distance downwind of a boundary, for fetch distances of 0.5 Ru, 2 Ru and 3 Ru the portion of relocated snow evaporating in transit would be about 37, 57, 75 and 83 percent respectively.

III.1.3 Combined Transport and Evaporation

The work by Tabler (1975) led to the development of a combined transport and evaporation equation of the form:

$$Q = \sum_{i=1}^{n} \left( P_{r_i} \right)_{i} \left( 0.14 \frac{R}{Ru} \right) dr$$

where $Q$ = the total water equivalent volume of snow arriving at a point after a transport distance, R, (m$^3$/m),

$n$ = number of interval distances of $\Delta R$ within which R can be divided over which $P_{r_i}$ and evaporation can be assumed constant,

$\left( P_{r_i} \right)_{i}$ = the water equivalent of the precipitation swept off the i th AR interval (m) - for annual transport calculations $P_{r_i}$ is the total winter precipitation less the amount held by the vegetation and terrain irregularities plus the amount of melt,

R = the travel distance, (m),

Ru = the transport distance required for complete evaporation of the average size particle (evaluated using measured field data) (m),

$dr$ = incremental length (m).

In his paper Tabler proceeds to demonstrate the application of the model for determining transport volumes and evaporation losses for different combinations of vegetation and terrain features.

In manner of summary, Miller (1976) lists the actual transport distances and the transport flux rates for different physiographic and climatic regions (see Tables 2 and 3). Steppuhn and Gray (1978) have estimated the potential transport fluxes over the winter season on the Canadian Prairies to vary in the range...
Table 2. Average distances of snow transport in different terrain, km, (after Miller, 1976).

<table>
<thead>
<tr>
<th>Terrain Description</th>
<th>Distance Range (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>In mountainous dissected topography (Kotlyakov, 1973)</td>
<td>0.1 to 0.5</td>
</tr>
<tr>
<td>On plains of western Siberia (Dyunin et al, 1973)</td>
<td>1 to 3</td>
</tr>
<tr>
<td>On plateaus of southeastern Wyoming (Tabler &amp; Schmidt, 1973)</td>
<td></td>
</tr>
<tr>
<td>---small particles (0.1 mm diam.)</td>
<td>0.5</td>
</tr>
<tr>
<td>---larger particles (0.2 mm diam.)</td>
<td>1.4</td>
</tr>
<tr>
<td>On ice domes (Kotliakov, 1968)</td>
<td>Up to 5</td>
</tr>
<tr>
<td>On polar ice caps (Kotliakov, 1968)</td>
<td>To 30-50</td>
</tr>
</tbody>
</table>

Table 3. Mean winter season transport flux rates, tons/m (after Miller, 1976).

<table>
<thead>
<tr>
<th>Location</th>
<th>Flux Rate Range (tons/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Central European Russia (Mikel', et al, 1969, p.54)</td>
<td>100 to 200</td>
</tr>
<tr>
<td>Kamennala Steppe (Mikel', et al, 1969, p.54)</td>
<td>230</td>
</tr>
<tr>
<td>Tundra (Mikel', et al, 1969, p.54)</td>
<td>600 to 1000</td>
</tr>
<tr>
<td>Arctic Coast (Mikel', et al, 1969, p.54)</td>
<td>1000</td>
</tr>
<tr>
<td>Wyoming (Tabler, 1973)</td>
<td>150</td>
</tr>
<tr>
<td>Rocky Mountains (Martinelli, 1973)</td>
<td>300</td>
</tr>
<tr>
<td>Northern Idaho ridge: Increase in transport rate into a lee cornice when windward slope was cleared (Haupt, 1973)</td>
<td>14 to 120</td>
</tr>
</tbody>
</table>

from 2.6 to 21.7 tonnes/m.

III.1.4 Interaction in Forest Environment

In the forest environment maximum accumulations of snow are often reported at the edges of a forest as a result of snow being blown in from adjacent areas. These accumulations are highly dependent on the porosity of the stand borders. Within the stand itself snow accumulations may not be uniform although it is generally conceded that the snow cover distribution is more uniform within hardwoods than in coniferous forests where the effect of dense crowns is to create a "ridge and hollow" effect. Further, most studies have reported that more snow is found within forest openings than within the stand itself.

A unique transport phenomenon affecting the distribution of snow cover within a forest which is affected primarily by wind is the transport of intercepted snow. Wind causes a tree to vibrate, thereby resulting in a loosening and erosion of intercepted snow, blowing and transporting fragments downwind. Miller (1966) in his review paper on the transport of intercepted snow summarizes the results of different studies concerning the transport process. It is obvious from this review that a physical understanding of the process is complicated not only by the complex nature of the airflow patterns and velocity distributions within different forest covers but also because of other factors such as the manner with which snow accumulates, collects and adheres to different types of vegetation and the cohesion and adhesion properties of different types of snow. Although several investigations have been conducted of the critical wind speed required to remove intercepted snow the writer is unaware of any field measurements which provide quantitative measures of the amount of transport by wind and its affect on snow cover distribution. Hoover and Leaf (1967) in Colorado did conclude from timed-sequence photographic measurements of snow interception that there was a possibility that mechanical removal predominated over vaporization in removing snow from tree crowns.

III.1.5 Effects on Snow Density

To the hydrologist a discussion of wind on a snow cover would be incomplete without some consideration of its affect on snow density and consequently the snow water equivalent. When snow crystals are moved by wind their physical shape and properties are changed and they are redeposited in accumulations which have densities greater than the parent materials. For example, Church (1941) found fresh snow with densities of 36 kg/m$^3$ and 56 kg/m$^3$ to increase to a density of 176 kg/m$^3$ within 24 hours after being subjected to wind action. Gray et al (1971) reported similar findings for the Prairies.
densities of snow as influenced by wind.

The time densification of a snowcover, although initiated by the wind action is also influenced by condensation, melting and other processes. These differences are well illustrated by considering the curves for different vegetation zones (see Figure 9).

III.2 Topographic Factors

III.2.1 Elevation

The primary topographic factors affecting snow accumulation and distribution are elevation, slope and aspect. Of these three normally, in a mountain terrain, elevation is considered the major factor and, at a specified location and within a given elevation interval frequently a linear association between snow accumulation and elevation can be found (see for example U.S. Corps of Engineers, 1956). The transposability of these relationships is highly suspect as the influence of elevation alone is indeterminate because of the interdependency of climate, slope and elevation.

Figure 8, constructed from data reported by Meiman (1970) shows the increase in snow accumulation with elevation as measured in eleven separate investigations undertaken in Alberta, California, Colorado, Idaho and New Mexico. In one study (Colorado) measurements were made within selected elevation bands for three consecutive years. The information reflects not only the large variation encountered between major physiographic areas as well as the spatial-temporal variations within a given area. Meiman

Table 4. Densities of Snow Cover

<table>
<thead>
<tr>
<th>Snow Type</th>
<th>Density (kg/m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wild snow</td>
<td>10 to 30</td>
</tr>
<tr>
<td>Ordinary new snow immediately after falling in the still air.</td>
<td>50 to 65</td>
</tr>
<tr>
<td>Settling snow</td>
<td>70 to 90</td>
</tr>
<tr>
<td>Very slightly toughened by wind immediately after falling</td>
<td>63 to 80</td>
</tr>
<tr>
<td>Average wind-toughened snow</td>
<td>280</td>
</tr>
<tr>
<td>Hard wind slab</td>
<td>350</td>
</tr>
<tr>
<td>New firn snow</td>
<td>400 to 550</td>
</tr>
<tr>
<td>Advanced firn snow</td>
<td>550 to 650</td>
</tr>
<tr>
<td>Thawing firn snow</td>
<td>600 to 700</td>
</tr>
</tbody>
</table>

*a* snow consolidated partly into ice (after Seligman, 1962)
cites that numerous workers have substantially improved the correlations by including other land surface features. However, as Peck (1964) and others have emphasized, and as the writer has alluded to in prior discussion it is important to recognize the influence of climatic factors or elements of the parent weather system in interpreting snow distribution and accumulation patterns. For example, the increase in temperature found with elevation results in a decrease in melt losses. Also, it can be argued that since the moisture available for the precipitation process decreases with elevation the increases observed with elevation reflect the combined influence of slope and elevation on the efficiency of the precipitation mechanism.

To the hydrologist, the density of a snow cover is equally important as depth. Storr and Golding (1974) show a linear association between snow water equivalent and depth in the Harmut Basin in Alberta. Grant and Rhea (1974) report a substantial variation of snow density in mountain passes in the central Colorado Rockies. Greater densities were observed at the lower elevations which was attributed to a higher frequency and degree of riming of ice crystals than occurs at higher elevations.

III.2.2 Slope

Mathematically, the orographic precipitation rate is predominately related to terrain slope and windflow rather than elevation. That is, if the air is saturated the rate at which precipitation is produced is directly proportional to the rate of rise of the air and the rate of rise of the air flowing over a upsloping terrain is directly proportional to the product of the windspeed and the magnitude of the slope.

Rhea and Grant (1974) have provided a general expression for orographic mountain snowfall to include the effects of large-scale vertical air mass movement, convective activity and orographic lift, passage of air mass over upwind mountain barriers, interception and gauge exposure. The orographic component is mathematically related to the specific humidity gradient, a vertical velocity component—a function of slope and time, the pressure depth of the air mass, the precipitation efficiency and the elevation of the site. They found from analysis of winter precipitation data in Colorado that the long-term average precipitation at a point was strongly correlated (positively) with topographic slope computed in the first 20 km upwind, using the slope values obtained from 30° directional classes and weighted according to the directional frequency in precipitation days and an estimated cloud depth for each direction. The correlation between measured winter precipitation and computed "orographic" precipitation was slightly improved (r=0.082 to r=0.90) by incorporating a factor to account for the partial depletion of available condensate because of precipitation induced by upstream barriers. They concluded that long term average precipitation is not well correlated to station elevation except for points on the same ridge.

Even where orography is the principle lifting mechanism and an increase in snowfall amounts with elevation may be expected the depth of accumulation or deposition may not exhibit this pattern. In addition to the many factors affecting distribution at the higher elevations winds of high velocities and long duration are more frequent causing transport and redistribution. Table 5 reported by Miller (1976) shows the high variability in the relative amounts of snow deposited on different topographic facets in the Ural Mountains in the USSR.

In topographically-similar areas to the Prairies in which snow occurs primarily by frontal activity and the exposed snowcover is subjected to high wind shear forces slope and aspect are important terrain variables affecting the snowcover distribution. The depth of snowcover along a slope oriented in the direction of the prevailing wind tends to decrease along its length. Steppuhn (1978) shows that the relative amounts of snow retained by level plains, gradual slopes and hill tops, all in summer fallow, to that measured by a Nipher snow gauge to be of the order of 0.6, 0.70 and 0.2 respectively (see Table 8). In many years on the Prairies it is common to note, even at the time of maximum accumulation, hilltops may be free of snow. Further, it should be noted that in this region the lee of steep slopes and gullies form major collection areas. McKay (1970) cites the case where the snow water equivalent remaining in a prairie valley in the spring after the adjacent plains were free of snow amounted to 14,000 m³ per km of channel. Steppuhn provides examples which suggest that the retention in small drainage-ways and steep hills and valley slopes compared to gradual slopes...
Table 5. Relative snow deposition on topographic facets along a traverse across the Ural Mountains (Kotliakov, 1968, p.161) (from Miller, 1976).

<table>
<thead>
<tr>
<th>Topographic Facet</th>
<th>Length, km</th>
<th>Precipitation (Relative)</th>
<th>Accumulation (Relative)</th>
<th>Eddy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flat</td>
<td>10.0</td>
<td>1.0</td>
<td>1.0</td>
<td></td>
</tr>
<tr>
<td>Lower slope</td>
<td>2.0</td>
<td>1.75</td>
<td>1.2</td>
<td></td>
</tr>
<tr>
<td>Upper slope</td>
<td>1.0</td>
<td>1.25</td>
<td>0.25</td>
<td></td>
</tr>
<tr>
<td>Ridge</td>
<td>1.3</td>
<td>1.4</td>
<td>2.1</td>
<td></td>
</tr>
<tr>
<td>Lee of main</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Second ridge</td>
<td>1.0</td>
<td>0.5</td>
<td>1.1</td>
<td></td>
</tr>
<tr>
<td>Lee of second ridge</td>
<td>0.8</td>
<td>0.6</td>
<td>0.6</td>
<td></td>
</tr>
<tr>
<td>Downwind plateau</td>
<td>0.7</td>
<td>0.8</td>
<td>0.9</td>
<td></td>
</tr>
<tr>
<td>Valley of Igan River</td>
<td>0.7</td>
<td>0.7</td>
<td>1.3</td>
<td></td>
</tr>
<tr>
<td>Windward slope</td>
<td>0.5</td>
<td>0.9</td>
<td>0.3</td>
<td></td>
</tr>
<tr>
<td>Ridge and its lee slope</td>
<td>1.3</td>
<td>0.6</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>Lower lee slope</td>
<td>0.5</td>
<td>1.0</td>
<td>0.9</td>
<td></td>
</tr>
</tbody>
</table>

Standard deviation: 0.43 in H; 0.52 in V

a H—Horizontal-axis eddy, V—Vertical axis eddy.

Table 6. Mean Accumulation of Snow (cm) under different cover conditions in the Eastern Rockies as related to aspect (after Stanton, 1966).

<table>
<thead>
<tr>
<th>Cover</th>
<th>Aspect</th>
<th>N</th>
<th>S</th>
<th>E</th>
</tr>
</thead>
<tbody>
<tr>
<td>Forest</td>
<td></td>
<td>41</td>
<td>41</td>
<td>39</td>
</tr>
<tr>
<td>Cut Forest</td>
<td></td>
<td>45</td>
<td>53</td>
<td>65</td>
</tr>
</tbody>
</table>

The importance of aspect with respect to producing major changes in climatic regimes and energy exchange processes is less apparent when local conditions are considered. Within the Prairie environment it is accepted that the influence of aspect on accumulation is dominated by its influence on the latter and, in addition, the snow transport phenomenon.

No attempt will be made within this manuscript to discuss the complex interactions between melt, ablation and accumulation and distribution. The melt and ablation phenomena are a separate topic of this Workshop.

III.3 Vegetative Cover

Vegetation affects the surface roughness and wind velocity thereby affecting the erosional, transport and depositional characteristics of the surface. If the biomass extends above the depth of the snowcover it affects the energy exchange processes, the magnitudes of the energy terms and the position (height) of the most active exchange surface. Also, a vegetative canopy affects the amount of snow reaching the ground. In most cases the works concerning the
interaction of vegetation and snow accumulation are divided to study of forest and non-forest (short vegetative cover) ecosystems.

III.3.1 Forest

A forest differs from other vegetative covers primarily in its protrusion of large intercepting and radiating biomass above the snowcover surface and into the atmosphere.

Kitteredge in 1953 showed using illustrations, the effects of a forest canopy on local snowcover distributions. Since that time numerous studies have been conducted on the canopy effects on interception and accumulation. These are summarized by Miller (1966) and Meiman (1970). Most of the works have concentrated on differences between the amounts of snowcover under different species of forests and the amounts measured in openings. Though more snow is consistently found in forest openings than within the stand itself absolute estimates of the interception amounts are unknown because of the lack of knowledge of the interception loss processes.

Kuzmin (1960) reported that the amount of snow accumulated in a forest clearing depends on the size of the wooded area. The average depth in clearings in a wooded area 100 m x 200 m was 53 cm compared to a depth of 36 cm in an area 1000 x 2000 m. He also reported that the accumulations depend on the type of tree species; deciduous forests giving greater accumulations than pine or fir forests; the latter showing accumulations of 82 and 63 percent of that measured in a birch forest. Multi-factor analyses have disclosed snowcover amounts to increase from 8 to 56 mm of water equivalent for a 10% decrease in canopy depending on tree species (Meiman, 1970).

Many investigators (Goodell, 1959; Molchanov, 1963) have found snow accumulation to be inversely related to canopy density. Hence, it may be expected that forest density can be used as an index of the amount of interception. Kuzmin (1960) reports that the relationship between the water equivalent in a fir forest ($W_{E_f}$) and the water equivalent of the snowcover in a clearing, ($W_{E_c}$), can be related to tree density ($p$), expressed as a fraction by the expression

$$W_{E_f} = W_{E_c}(1-0.37p) \quad (6)$$

In addition to the effects of a forest cover on the wind velocity distribution and interception the other major effect of cover on snow accumulation and distribution originates from its effect on the energy exchange processes and thereby the way in which the snowcover is modified by the energy fluxes to change its erodability, mass and state. The classical work of the U.S. Corps of Engineers (1956) and Reifsnyder and Lull (1965) serve as excellent guides to the effects of a forest canopy on the radiative exchange components. More recently Bohren (1972) presented a theoretical analysis of radiation transfer and snowcover and has suggested that the total surface area of the canopy is a relevant parameter which determines the amount of radiant energy incident on a snowcover. The turbulent transfer of the sensible and latent heat between snowcover and the atmosphere in a forest are largely unknown quantities.

In manner of summary it is suggested that at the present time because of the lack of understanding of snow transport processes and the lack of information on the magnitudes of radiative and turbulent fluxes the accretion of snowcover in a forest is indeterminate. Miller (1966) summarized the types of research studies needed of heat and mass transfer quantities to obtain an understanding of the interception and transport phenomena in forests. This work serves a useful guideline to the development of an accumulation model.

III.3.2 Prairies (Grassland) and Steppes

Kuzmin (1960) and McKay (1963) have emphasized the importance of terrain and wind in establishing snowcover patterns on the Prairies. However, over the highly-exposed, relatively-flat or moderately-undulating terrain, meso- and micro-scale differences in vegetation as it increases aerodynamic roughness may produce wide variability in accumulation patterns. Accumulations are most pronounced where there is a long upstream fetch of loose snow which is subjected to sustained strong winds from one direction; they are less pronounced when winds change direction especially if the wind velocities are low.

Lakshman (1973) cites examples where on areas stratified according to land use and topography, a linear relationship may be used to relate snow water equivalent and snow depth. Steppuhn and Dyck (1974) suggest consistent similarities exist in
the areal variation of snowcovers within areal units having similar landscape features. That is, forests, pastures, cultivated fields, ponds, etc., within the same climatic region tend to accumulate snow according to recurring patterns unique to specific terrain and land use. The authors demonstrate the value in using a stratified or unitized sampling technique in estimating true basin snowcover. The procedure is based on the division of a watershed into different landscape classes and sampling within given areal units of a given class. Table 7, taken from Steppuhn (1976) shows the snowcover depth statistics by landscape type measured on the Creighton Tributary of the Bad Lake Watershed located in west central Saskatchewan. It is obvious from these data

Table 7. Snowcover depth statistics by landscape type, 1974, Bad Lake, Saskatchewan (after Steppuhn, 1976)

<table>
<thead>
<tr>
<th>Landscape Type</th>
<th>No. Obs.</th>
<th>Mean Depth (cm)</th>
<th>Coeff. of Variation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plain fallow</td>
<td>360</td>
<td>41.5</td>
<td>0.155</td>
</tr>
<tr>
<td>stubble</td>
<td>216</td>
<td>46.4</td>
<td>0.133</td>
</tr>
<tr>
<td>Rolling Plain fallow</td>
<td>668</td>
<td>49.4</td>
<td>0.151</td>
</tr>
<tr>
<td>stubble</td>
<td>180</td>
<td>58.8</td>
<td>0.083</td>
</tr>
<tr>
<td>pasture</td>
<td>578</td>
<td>56.2</td>
<td>0.174</td>
</tr>
<tr>
<td>Gradual Slope fallow</td>
<td>324</td>
<td>50.4</td>
<td>0.202</td>
</tr>
<tr>
<td>stubble</td>
<td>180</td>
<td>47.4</td>
<td>0.147</td>
</tr>
<tr>
<td>scrub</td>
<td>183</td>
<td>65.6</td>
<td>0.240</td>
</tr>
<tr>
<td>Slough fallow</td>
<td>180</td>
<td>46.1</td>
<td>0.110</td>
</tr>
<tr>
<td>stubble</td>
<td>108</td>
<td>46.6</td>
<td>0.139</td>
</tr>
<tr>
<td>Sharp Slope pasture</td>
<td>400</td>
<td>111.5</td>
<td>0.199</td>
</tr>
<tr>
<td>scrub</td>
<td>869</td>
<td>126.5</td>
<td>0.239</td>
</tr>
<tr>
<td>Broad Lowland fallow</td>
<td>395</td>
<td>101.1</td>
<td>0.188</td>
</tr>
<tr>
<td>stubble</td>
<td>435</td>
<td>95.2</td>
<td>0.084</td>
</tr>
<tr>
<td>pasture</td>
<td>219</td>
<td>97.0</td>
<td>0.114</td>
</tr>
<tr>
<td>scrub</td>
<td>537</td>
<td>112.3</td>
<td>0.183</td>
</tr>
<tr>
<td>Topland fallow</td>
<td>507</td>
<td>22.9</td>
<td>0.277</td>
</tr>
<tr>
<td>stubble</td>
<td>218</td>
<td>37.6</td>
<td>0.136</td>
</tr>
<tr>
<td>pasture</td>
<td>181</td>
<td>21.2</td>
<td>0.434</td>
</tr>
<tr>
<td>Farm Yard fallow</td>
<td>72</td>
<td>129.1</td>
<td>0.182</td>
</tr>
</tbody>
</table>

that the depth of snow collected by scrub brush is consistently higher than that collected on fallow, stubble or pasture independent of terrain. Conversely, the strong interdependency of vegetation and terrain in relation to the comparative amounts of snow retained by fallow, stubble and pasture can be noted. It is worthy to note that 1973-1974 was a winter with snowfall much above normal, the mean snow water equivalent on the basin was about 134 mm. Using data from 1973/74 and 1974/75, the latter a winter in which the mean areal snow water equivalent was measured as 71.2 mm Steppuhn (1978) normalized the mean water equivalent measured on the different areal units to that measured by a Nipher gauge, (see Table 8). It can be noted that (1) the ratios for a given landscape type differ in the two years, (2) within the different landscape classes the ratios remain relatively consistent in order of magnitude with respect to vegetation types, and (3) there is a reasonable degree of order with which the different landscape types accumulate snow; the accumulations on steep hill and valley slopes tending to be much greater relative to those of other landscape types in the year of lower snowfall. It should be recognized that these data are presented for example purposes only and are limited in their application as they represent only the results of two years of study.

### III.3.3 Zonal Snowcover

The discussions indicate that climate physiography and vegetation interact in a complex manner in governing snowcover accumulation and distribution. Many investigators (Richter, 1945; Khodakov, 1975; and McKay and Findlay, 1971) have found that it is possible to establish and to map some general characteristics of snowcover on a zonal basis in which the zones are defined according to vegetation. This procedure is often effective because the type of vegetation is frequently indicative of climate. Usually this simple zonation includes the following: Tundra, Taiga and Boreal Forest, Grassland and Steppes, Mixed Forest and Mountain Areas. It can be expected that between the different zones, and in fact within zones, the data of formation, duration, date of disappearance, depth and density of the snowcover vary widely depending on such factors as latitude, altitude and other physiographic features. Nevertheless, independent of the wide differences in the physical properties of the snowcover within the
Table 8. Ratios of mean snow water equivalent on given landscape class to snow water equivalent measured in a Niipher gauge - Creighton Tributary, Bad Lake Watershed, Saskatchewan, 1974 and 1975.

<table>
<thead>
<tr>
<th>Landscape Type</th>
<th>Ratio of mean snow water equivalent on landscape type to that measured by Niipher gauge.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1974a</td>
</tr>
<tr>
<td>Upland and Lowland Plains (shallow sloughs)</td>
<td></td>
</tr>
<tr>
<td>fallow</td>
<td>0.59</td>
</tr>
<tr>
<td>stubble</td>
<td>0.68</td>
</tr>
<tr>
<td>hayland, pasture</td>
<td>0.59</td>
</tr>
<tr>
<td>Undulating Plains</td>
<td></td>
</tr>
<tr>
<td>fallow</td>
<td>0.78</td>
</tr>
<tr>
<td>stubble</td>
<td>0.78</td>
</tr>
<tr>
<td>hayland, pasture</td>
<td>0.78</td>
</tr>
<tr>
<td>Gradual Hill and Valley Slopes</td>
<td></td>
</tr>
<tr>
<td>fallow</td>
<td>0.70</td>
</tr>
<tr>
<td>stubble</td>
<td>0.73</td>
</tr>
<tr>
<td>hayland</td>
<td>0.73</td>
</tr>
<tr>
<td>hayland, pasture</td>
<td>0.94</td>
</tr>
<tr>
<td>Steep Hill and Valley Slopes</td>
<td></td>
</tr>
<tr>
<td>fallow</td>
<td>2.06</td>
</tr>
<tr>
<td>stubble</td>
<td>1.99</td>
</tr>
<tr>
<td>hayland, pasture</td>
<td>1.99</td>
</tr>
<tr>
<td>bush</td>
<td>1.99</td>
</tr>
<tr>
<td>Small Shallow Drainageways</td>
<td></td>
</tr>
<tr>
<td>fallow</td>
<td>1.67</td>
</tr>
<tr>
<td>stubble</td>
<td>1.67</td>
</tr>
<tr>
<td>hayland, pasture</td>
<td>1.67</td>
</tr>
<tr>
<td>scattered brush</td>
<td>1.67</td>
</tr>
<tr>
<td>Wide Valley Terraces and Bottoms (ravines, coulies)</td>
<td></td>
</tr>
<tr>
<td>fallow</td>
<td>0.74</td>
</tr>
<tr>
<td>stubble</td>
<td>0.75</td>
</tr>
<tr>
<td>hayland, pasture</td>
<td>0.75</td>
</tr>
<tr>
<td>brush</td>
<td>0.75</td>
</tr>
<tr>
<td>Farm Yards</td>
<td>2.10</td>
</tr>
</tbody>
</table>

a) Ratio of areal mean water equivalent to Niipher gauge: 1974-0.729 1975-0.82

different zones it is also possible to distinguish major differences in certain properties between zones. For example, Figure 9 exhibits the differences in the time-density variations in different vegetation regions. Although such information can be used by a hydrologist only for grossly simplified computations of water equivalent they serve the purpose of illustrating the effects of vegetation on the ripening process under different conditions.

IV SIMULATION, SYNTHESIS AND PREDICTION

The snow transport and accumulation phenomena are highly complex and the exact mechanisms of movement are difficult to explain even under the simplest situation of a smooth flat infinite plane. The following section discusses some of the physical, mathematical and empirical, to include measurement and conceptual approaches, used to describe the areal accumulation of snowcover.

The approach adopted to study and to determine snowcover distribution will depend largely on such factors as: the study objectives, the available resources and the use of the results. That is, an engineer whose primary interest is to study distribution patterns surrounding buildings, air terminals and runways, roadways and other engineering works resulting from a single storm may take advantage of the use of physical models. Conversely, to a hydrologist, whose interest is in the areal snowcover distribution, primary snow water equivalent, at the time of maximum accumulation the use of a physical
model is completely impractical not only because of its physical size and cost but more so because most natural systems defy the construction of a model that will satisfy the geometric, kinematic and dynamic similitude criteria.

To the hydrologists the most important uses of snowcover data are for forecasting, or updating forecasts of water yield or the magnitude and time of occurrence of flood peaks. In many climatic environments absolute data of areal snowcover distribution, by itself, will not necessarily lead to significant improvements in forecasts than can be obtained through the use of one-dimensional snowmelt models currently available. The point to be emphasized is that the areal distribution of snowcover and snow water equivalent does not necessarily accurately reflect the major source areas of snowmelt runoff. Snowmelt, prior to becoming streamflow, is modified by ablation (snowmelt and infiltration) and storage processes. Hence it is only where the boundary conditions affecting these other processes remain sensibly constant, both spatially and temporally, or where the effects or interactions between the snowcover and these processes can be explained will it be possible to make efficient quantitative use of areal snowcover differentiation in streamflow prediction. To exemplify, Male (1972) in a study of the thermodynamics of the melt of a shallow snowcover at a point identified that the depth and density (snow water equivalent) are the most critical properties affecting the changes in internal energy of a shallow snowcover and the sensitivity of the energy budget approach for determining snowmelt. He exemplified that the cumulative uncertainty to be placed on estimates of the internal energy may be several multiples (ranging between 2 to 16), of both the net radiation and sensible heat fluxes depending on the magnitudes of the different components; the larger the fluxes the smaller the ratio. The results of this study define the importance of the depth and density of snow in relation to the time of occurrence of melt and the amount produced. Erickson et al (1978) reported the results of a study conducted on small units on the Prairies which show that the runoff volumes and rates are greatly affected by land use and to a lesser extent terrain. Table 9, taken from these works illustrates the volumetric differences which may be found under differing land use or vegetal covers based on the degree day approach for estimating the runoff quantities. The data indicate: (a) the differences in unit runoff volumes in forest and Prairie regions - the runoff volumes per degree day, in all cases but one (stubble) exceeding those experienced in a forest and (b) the effects of land use on runoff within the Prairies - the average values being 3.90 for stubble, 4.78 for pasture and 10.96 for fallow (units in mm of water per °C-day above 0°C) with maximum rates for the three land use conditions reported as 5.04 6.50 and 18.62 respectively. These rates are in reverse order of magnitude relative to the depth of snow accumulation on the different land use classes; the depth on pasture and stubble being significantly greater than on fallow. Many factors have contributed to the differences in the release rates and a comprehensive discussion of each is beyond the scope of this paper. The important aspects of the study are that on the Prairies the magnitudes of rates and volumes of runoff from different land units may be in complete opposition to the snowcover depth distribution patterns.

McKay (1970) has pointed out that in many years on the Prairies the snow accumulated in gullies and drainageways (major accumulation areas) serve as major source areas of runoff. In addition these accumulations may retain large quantities of water which originate from snowmelt runoff from adjacent stubble, pasture and fallow land and delay the time of
Table 9. Comparison of basin snowmelt runoff degree day factors between forest and prairie environments.

<table>
<thead>
<tr>
<th>Location</th>
<th>Land Use</th>
<th>Month</th>
<th>Runoff/°C day above 0°C/unit snow covered area (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Central Sierra (Calif.)*</td>
<td>forest</td>
<td>April</td>
<td>4.07</td>
</tr>
<tr>
<td>Upper Columbia (Mon.)*</td>
<td>forest</td>
<td>April</td>
<td>1.69</td>
</tr>
<tr>
<td>Willamette Basin (Ore.)*</td>
<td>forest</td>
<td>April</td>
<td>.78</td>
</tr>
<tr>
<td>Bad Lake (Sask.)</td>
<td>stubble</td>
<td>Mar., Apr.</td>
<td>3.91</td>
</tr>
<tr>
<td>Bad Lake (Sask.)</td>
<td>pasture</td>
<td>Mar., Apr.</td>
<td>4.79</td>
</tr>
<tr>
<td>Bad Lake (Sask.)</td>
<td>fallow</td>
<td>Mar., Apr.</td>
<td>10.96</td>
</tr>
</tbody>
</table>

* Taken from U.S. Corps of Engineering (1956).

release to streamflow. Field observations have substantiated that measurable streamflow and floods will occur only after water has channelled the snow in the drainage ways; a situation that may occur when a large part of the snowcover has been removed from adjacent areas. To date, procedures for routing of snowmelt through snow-clogged channels to account for the storage effects have not been developed.

The preceding discussion emphasizes the problems of applying a snowcover distribution model to assist forecasting and of water yield and floods in the Prairie environment; the time elements of any release pattern may be directly related to snowcover depth whereas the volumetric elements may be indirectly related to depth. Certainly, in a mountain, forested region a similar dilemma exists independent of the fact that different factors may assume greater importance in the accumulation-ablation-runoff interaction than those in the Prairies.

It appears possible, however, that for hydrological purposes a snow accumulation and distribution model may be developed giving consideration to soil-vegetation-topography factors. Many investigators (Adams, 1976; Leaf, 1975; Steppuhn and Dyck, 1974 and operational agencies in the USSR) are proponents of this practice for application to rolling forest, mountain forest and prairie areas. The number of units or classes selected will depend not only on the areal variability of the different physical variables but also on such factors as the sensitivity of the given water management model to the variations in the input data, the type and availability of data acquisition, handling and processing facilities and the needs and requirements of the user.

IV.1 Physical Models (Kind)

Mellor (1965) has pointed out that the value of wind tunnel testing in snow drifting studies has been widely recognized for about 40 years although relatively few studies have been conducted during that time. By using sawdust and flake mica to simulate snow in a wind tunnel, Finney (1939) was able to suggest improved methods for snowdrift control along highways. More recently Theakston (1962) has studied snow accumulation around structures first by the use of wind tunnels and later by use of water flumes. Quite extensive studies on drifting snow have been conducted at New York University by Gerdel and Strom (1961). In the early tests modelling criteria were apparently not explicitly considered. It is entirely impracticable to satisfy the large number of similarity criteria yielded by a simple dimensional analysis of the system. However, a similarity analysis which draws more heavily on physical analysis and experimental evidence promises to be more fruitful. Strom et al (1962), Odar (1962, 1965), Isyumov (1971) and Kind (1976) have considered this problem.

Kind (1976) has presented three design criteria which must be satisfied to correctly model the ratios drag force: weight, drag force:inertia force and velocity fields in studying the free-flight portion of the trajectory of a saltating particle (the saltation process). A requirement for similarity of model and prototype velocity fields is that the model be geometrically similar to the prototype. In addition, the velocity profile of the wind approaching the model must be similar to that found in the prototype. Thus, the model profile must give the same value of \( U^* / V \) (where \( U^* \) is the shear velocity and \( V \) is the wind velocity) as the prototype profile. The model must also include a substantial fetch of correctly modelled terrain, complete with particles, upstream of the
region of interest. This is necessary in order that saltation be properly established by the time the flow reaches the region of interest. The requirement that \( V' / L g \) (where \( V \) is a reference wind velocity, \( L \) is a reference dimension and \( g \) is the gravitational constant) be the same in model and prototype will usually conflict that \( U*^2 / 2gV \) (where \( U* \) is the threshold shear velocity and \( V \) is the kinematic viscosity) be greater than 30 - a condition required for the flow over rough surfaces to be independent of the kinematic viscosity of the air.

Kind also discusses methods of selecting model particles. Of special interest is the fact that the mass density of the 'snow' particles will usually be different in the model and prototype cases and this must be taken into account when determining the times for volume accumulations of snow from the model tests.

The requirements for correct simulation of the free-flight of falling particles are included in the overall simulation requirements. Snowfall and snow drifting can therefore be modelled simultaneously without additional difficulty. Simulation of the snowfall process is accomplished by dropping particles into the flow at an appropriate rate from the roof of the wind tunnel.

Sometimes only the simulation of airborne blowing snow is necessary where ground-drift is irrelevant. If the airborne snow particles of interest are very small they essentially follow the air flow and smoke or dye tracers could be used instead of particles.

The ratio of drag force to lift force on snow particles is largely due to the large ratio of particle density. This would not be true if water were used as the model fluid and therefore the use of water to model ground drifting of snow is open to some question. Nevertheless, model studies in water flumes have been and will continue to be useful (Theakston, 1962; Isyumov, 1971). The flow of water over buildings and other surface features is certainly basically the same as that of air. The saltation process is also qualitatively the same in both fluids although there are substantial differences in magnitudes for certain effects. Therefore drift patterns obtained in a water flume should be reasonably realistic if the velocity profile in the flow approaching the model is adequately simulated and if the model itself is geometrically similar to the prototype. Such model tests would not, however, be expected to yield reliable quantitative results for rates of snow accumulation or drift depths. Results for drift shape and location also tend to be somewhat inaccurate in water flumes although they are usually qualitatively correct.

IV.2 Mathematical, Empirical and Other Approaches

The complex nature of the transport and deposition processes and the large number of factors affecting these processes defies the development of a generalized, physically-based mathematical model for describing areal snowcover accumulation and distribution. However, the works of Tabler (1975) on a combined transport and evaporation model and Rhea and Grant (1974) on a method for accounting for the effects of the orographic lift component on snowfall represent creditable contributions in providing a physical base for a better understanding of the accumulation process. It is recognized that such models involve a number of simplifying assumptions and they require field input data. As is frequently the case in hydrological studies many of the simplifications used in a theoretical analysis cannot be verified because of the lack of field data. Likely, this deficiency imposes one of the greatest constraints on future model developments.

If sufficient point field data have been obtained over an area, isopleths of such variables as depth, density or water equivalent can be drawn and these maps used as an estimate of the distribution of the snowcover parameter. Such maps provide gross simplifications of snowcover distribution yet when complemented with data obtained from aerial photographs or satellites serve a useful operational function. Another method of using point data to obtain basin averages is the standard Thiessen polygon approach.

Lacking an understanding of the physical processes governing snowcover distribution, of necessity, hydrologists have opted to the use of empirical relationships to define the spatial distributions of snowcover properties. As it would be expected, based on the material presented in earlier parts of this paper, these relationships make use of the association between the snowcover property and topographical, vegetative and land use features. It is reiterated that such relationships generally invoke the concept that within the same climatic region consistent similarities may be
found in the areal variation of snowcover within areal units having common landscape features. The extrapolation use and application of these empirical associations outside of the region in which they were developed is highly questionable. In many cases, the associations are established through standard regression techniques. The works of Meiman (1970) and the U.S. Corps of Engineers (1956) provide an excellent summary of the relationships developed between snowcover depth and elevation, slope, aspect and other factors in forested mountain regions. Much work in these regions has been directed to study of the effects of clearing size on accumulation. In the less densely forested regions of rolling topography of Eastern Canada, Adams and Rogerson (1968) and Adams (1976) focussed their work in study of differences in snowcover character in different vegetation zones. On the prairies, Steppuhn and Dyck (1974) demonstrated the value of sampling within areal units of watershed having similar land use, terrain and vegetal cover features to obtain an accurate estimate of the basin mean water equivalent (see Table 7). Broader scale studies (McKay, 1972 and Billelo, 1969) demonstrate differences in certain snowcover properties that may be expected in different climate and vegetation regions (see Figure 9). Any review of literature concerning the interaction of snowcover accumulation patterns and vegetation and topography would be incomplete without citing the work of Kuz'min (1960) containing the much cited table of "snow retention coefficients" for a range of landscape classes.

The observation that snow accumulation patterns may be explained in terms of landscape features serves as a useful tool toward gaining an insight of areal snowcover. The relative amounts accumulated by the different units may differ appreciably from year to year because of the variability in the aerodynamic processes affecting accumulation and, in addition, changes in land use. The transport-condensation model proposed by Tabler explained in an earlier section of this paper delineates some of these factors. In addition, as Adams (1976) points out, significant differences can be expected, for example in the maximum depths of snowcovers on selected vegetation types in individual years, and that the nature of the differences between strata (vegetative types) may not be consistent between years. He points out that the consistency of the differences between the depth, density and water equivalent of different vegetative-based snowcovers will depend largely on the consistency in the evolution of the snowcover from initiation to peak accumulation. In a region where a snowcover is subject to periods of melt or to rain the changes to its physical properties caused by these factors may mask the differences caused by vegetation. His findings would suggest that a colder environment is more conducive to the preservation of different depositional conditions on different landscapes.

In principle, the application of landscape-based classes to differentiate areal snowcover distribution or to determine the mean basin water equivalent is relatively straightforward. Through the use of areal photographs and vegetative maps the hydrologist identifies, interprets, geographically locates and determines the areas of the different landscape classes. Selected units of the different classes are sampled to obtain statistically valid estimates of the mean depth, density or water equivalent for each class. These values are used for mapping and calculations. In many cases the procedure may not be accepted for operational purposes by management agencies:

(1) The division of a basin to different landscape classes requires prior knowledge and experience of snow accumulation facets and the preparation of maps is a time consuming task, independent of whether computer mapping facilities are available. New maps may need to be prepared each year depending on land use or vegetal cover changes caused, for example, by farming practices, forest manipulations, fires or other natural or artificial practices.

(2) The method requires comprehensive snowcover data from representative classes to establish reliable statistical estimates of the snowcover properties.

(3) Provision must be made for updating the snowcover conditions to account for precipitation, sublimation/condensation and melt following the date of the last survey.

Members of the Division of Hydrology, University of Saskatchewan are currently investigating the application of the areal differentiation of snowcover by landscape classes for prairie watersheds, in areas not subject to chinook activity, major winter sublimation losses, or melt. Attempts are being made to use Landsat imagery obtained in the fall to identify broad land use classes (ie. fallow, stubble,
sloughs and lakes, etc.). The advantages in use of these data are that they are in computer compatible form, a pixel gives a linear resolution of approximately 50-80 m (at ~ 50°N) thereby providing relatively-fine-scale areal delineation, the geographical position of each pixel is defined and the large number of brightness levels of the pixel provide flexibility and alternatives of investigating different features. The decision to apply satellite imagery in this manner rather than to attempt to use it directly to measure areal snow water equivalent was based on the uncertainty that the imagery would provide adequate estimates of water equivalent and, because of the orbital resolution, the uncertainty of obtaining quality imagery at the time of maximum accumulation. In the fall, prior to the occurrence of permanent snowcover and following harvest several passes of the satellite are made which usually provide one or more images under clear sky conditions. Terrain data will be stored vide one or more images under clear sky conditions. Terrain data will be stored.

The magnitudes of the coefficients c and d are mean values that describe the persistence in the local patterns of snow distribution on the glacier. Values of the coefficients a and b are established each year from depth and density samples taken at selected sampling points in different altitude zones. Using a linear relationship between depth and water equivalent with altitude the values are applied to all points on the grid to determine an estimate of the snow water equivalent for the entire glacier. The snowcover is then redistributed according to the equations while keeping the total amount of water constant. Maps are prepared from the adjusted values. A standardized data bank structure and the use of high speed data reduction systems make the method suitable for real time melt prediction studies.

It is appropriate to mention that the use of the grid square technique has been popularized for use in the spatial extrapolation of hydrometeorological data because of its adaptability for computer operations. Solomon et al (1968) for example, has applied the procedure in estimating precipitation, temperature and runoff for Newfoundland, Canada.

Even in the use of the more sophisticated mathematical models with distributed parameters, such as those used in streamflow forecasting, it is evident that the snowcover input values will be spatially-averaged either by smoothing or dividing a watershed to specified landscape units. Stratification, simply by means of providing improved areal representativeness reduces the systematic errors in estimating areal snowcover. Likewise the technique reduces the interpolation error within a physiographic region and provides a basis for statistical estimates of the same. Steppuhn and Dyck (1974) and Woo and Marsh (1977) have substantiated that significant improvements in the calculation of the areal mean are obtained.

Any discussion of the use of landscape-course snow surveys would be incomplete without recognizing that the procedure is in operational use in the USSR. Uryvaev and Vershinina (1971) report that

The depth of the snow surface is defined by the equation:

\[ d = a + bz + cR + dS \]  

where \( d \) = the depth of snowcover,  
\( z \) = the altitude of the point,  
\( R \) = the local relief,  
\( S \) = the surface slope angle, and  
\( a, b, c, d \) = empirically-derived coefficients.
landscape course snow surveying can be used to determine the water equivalent over areas of 250-1000 km$^3$ with an error of 5-10% compared with the results from continuous, comprehensive snow surveys.

Vershinina (1971) presents a comprehensive review, which includes both theoretical considerations and field results of the accuracy in determining the water equivalent of snowcover at a point, over a course and the average water equivalent of an area (based on landscape snow courses). His findings indicate an exponential type decrease in the root mean square error with increasing area for a constant network density; the magnitude of the error increases with a decrease in density. Some other important findings of the study are:

1. Errors in determining the snow water equivalent of an area are approximately the same for fields and forests for any network density.

2. The accuracy in determining the water equivalent of snow over an area with small absolute snowcover and during winter thaws is very low and depends markedly on the size of the area and network density. The errors under these conditions may be 50-100% higher than those calculated during the time of maximum water equivalent on the area.

3. The relative errors during maximum accumulation may be of the order of 15 to 20% in highly exposed areas where a shallow snowcover may be subject to frequent drifting.

4. The errors in determining snow water equivalent of an area are close to those for rainfall with a given network density.

5. The magnitudes of the errors for areas of equal size having the same network density exhibit regional differences. In a later paper Chermerenko (1975) substantiated these findings. His work extends that of Vershinina by including the effects of gauge location on network configuration on the magnitude of the error.

IV.3 Remote Sensing of Snowcover

(Goodison and Ferguson)

Ground based techniques of snowcover measurement will not always detect significant areal variations in snowcover. In mountain regions large variations in snow depth and water equivalent occur because of variations in slope, aspect, elevation, exposure and surface cover. In some areas, accessibility can limit both the number and representativeness of ground measurements. In plains regions, variations in snowcover are dominated by local land use and topographic variations, in addition to initial spatial variations in snowfall.

For hydrologic analyses, the percentage snowcover in a basin is an important variable. It is an areal parameter which ground measurements alone may not provide with sufficient accuracy, especially in sparsely-instrumented regions. However, the rapid development during the last 20 years of remote sensing techniques as applied to snowcover has provided new methods for observations, measurement and analysis. The successful application of these methods depends on accurate "ground truth" data for calibration and the verification.

IV.3.1 Aircraft Observations of Snowcover

With improvements in both photographic methods and airplane technology during the past 60 years, aerial photography has developed into a major method of photographing, interpreting, and mapping topography and other landscape features. However, to take vertical imagery over a large area is both time consuming and expensive.

The most common use of air photos is for topographic mapping; photography being done in late spring after the snowcover was gone, but before the leaves were on the trees. Such photography is of no value in the analysis of snowcover. Instead, special flights are usually required to obtain snowcover data. However, the rapid development of satellite remote sensing provides a promising alternative for the areal analysis of snowcover over large regions. Aircraft observations can then be used to obtain supplementary large scale data over selected areas of interest.

IV.3.2 Aerial Markers, Snowline Flights

In mountainous or remote regions where adequate snow course networks may be very expensive and difficult to service, and hence impractical, observations from aircraft may be a reasonable alternative. Snow depth at selected locations can be obtained by aircraft observations of previously installed aerial markers. An estimate of water equivalent for these sites is possible by using density information from nearby (within 40 km) snow courses located at similar elevations. Canadian and US agencies use daily runoff simulation models to predict runoff for
selected mountainous drainage basins in which an important parameter in the model is the snowline elevation that is, the boundary between the lower snow-free area and the higher snow-covered area. This elevation, in conjunction with the hypsometric curve for the drainage basin is used to calculate the percent of snow covered area for the basin. Field observations of the snowline from aircraft in British Columbia have been shown to greatly improve the prediction of a runoff simulation model (Sporns, 1976). For the Columbia River Basin both the United States Army Corps of Engineers and British Columbia Hydro and Power Authority conduct "snowline flights" to determine the snowline elevation. Personnel, flying in small aircraft at the altitude of lowest snowcover, estimate the snowline elevation for different aspects at preselected points marked on a topographical map. Two to four flights are made each year for each area being studied, May and June being the most common months because the snowline is most distinct.

Snowline flights are effective; yet are relatively expensive and time consuming. They may present a safety problem, and as well, a reliability problem as they are subject to suitable flying conditions, and the accuracy is somewhat dependent on the experience of the observer. As an alternative, efforts are being made to abstract snowlines from satellite imagery. Not only can a large area be mapped, but also more frequent observations are possible.

IV.3.3 Natural Gamma Radiation: Airborne Survey

The aerial gamma survey was developed in the USSR in the 1960's (Kogan et al, 1965) and since then experimental aerial surveys to measure snowpack water equivalent have been done in Norway, the USA and Canada.

Natural gamma radiation is emitted from the ground by potassium-40, bismuth-214 and thallium-206. The gamma radiation is attenuated by the mass of material between the source and the detector. Water, in either the solid or liquid phase, will attenuate gamma rays emitted from the soil surface. If the amount of attenuation is known, the amount of water causing the reduction can be calculated. The basic theory of the method and processing procedures of Canadian airborne measurements are summarized by Lojens and Grasty (1973). For snow water equivalent measurement, the potassium and integral (total count) data have been shown to be the most useful (Peck et al, 1971; Grasty et al, 1974; Jones et al, 1974).

As with the terrestrial survey, a no-snow measurement of the gamma levels along the traverse is necessary and adjustments for changes in soil moisture in the top 15 cm of the ground is required. The aerial survey method presents additional problems for which corrections must be made. The mass of air between the aircraft and the ground will attenuate gamma rays. Based on the temperature and pressure of the air at the flying height (typically 150 m), a correction for air mass attenuation can be calculated. Radon gas in the atmosphere also emits gamma radiation and it particularly affects those computations using the total count data. Entrapment of radon gas by an inversion layer near the surface can cause measurement problems since the background radiation will change. Background radiation is most easily measured by flying over a lake, since the mass of water will attenuate all gamma radiation emitted by the underlying ground. The reading over the lake establishes the zero value for the survey. If this method for establishing the zero value is not possible, other methods can be used, but with greater uncertainty (Jones et al, 1974). Finally, deviation from the pre-snow flight track, especially in areas of variable ground radioactivity, may introduce significant error in the snow water equivalent calculation. Variation in the count rate associated with changes in the background radiation, the mass of air below the aircraft, and the counting statistics of the equipment may introduce an error up to twice that associated with soil moisture variability.

An experimental gamma-ray spectrometer survey of snowcover over Southern Ontario gave encouraging results. The survey was done with a highly sensitive gamma spectrometer normally used for uranium surveys. The results indicated that the average snowcover water equivalent over 16 km sections could be measured to a precision of 12 mm using the potassium count and to 17 mm using the total count (Grasty et al, 1974). In this experiment, comparative ground based information was obtained from snow surveys along the flight track; snow water equivalents up to 140 mm were sampled.

The United States National Weather Service has conducted an aerial gamma survey over the Souris River Basin in the
central plains area of Canada and the USA at the time of peak snow accumulation. The aim of the survey is to provide near real time snow water equivalent data as an aid in snowmelt forecasting. Good results have been obtained; the estimated errors are of the same order of magnitude as found in the Canadian experiment (Loijens, 1975).

In contrast to snow course measurements, the gamma survey measures the mean water equivalent over an area. When used with fixed wing aircraft the requirement for low level flight restricts the technique to relatively flat areas; helicopters can be used in mountainous terrain. Precise navigation is required for duplicating tracks and comparing results to ground truth data. The great advantage of the aerial gamma technique is its coverage of large areas. By averaging the measured water equivalent in sections along the line the problem of high local variation in snowpack water equivalent, which commonly affects snow survey measurements, can be minimized. The present status of the gamma method indicates that in some cases it can be operationally feasible for obtaining the snowpack water equivalent over a basin.

However, as pointed out by Steppuhn (1976) sensing the snowcover attenuation of terrestrial gamma radiation may ease sampling, but will not eliminate it. He points out that in certain cases the extent of area sampled is relatively small in relation to the basin size. For example, the aerial gamma survey over the Souris River Basin covered only 0.42% of the watershed. The point made is that inherently the method suggests a grossly inflated sampling density. In most cases the accuracy with which the method will provide an estimate of the true basin water equivalent will depend directly on the representativeness of areas sampled.

IV.3.4 Microwave Sensing of Snowcover from Aircraft

Both radar and passive microwave systems operate in the portion of the electromagnetic spectrum from 0.1 cm to 100 cm. Passive microwave systems sense a portion of the natural radiation emitted by objects; radar is an active system which emits a beam of radiation and measures the reflected and back-scattered return. For snowcover applications both techniques are still in the research and development stage. Microwave emission (or brightness temperature) has been shown to change with snowpack accumulation and with the wetness of the snowpack (Meier, 1975). The great advantage with microwave systems is their day-or-night operational capability and their ability to sense through a cloud cover. Studies are currently being conducted on the usefulness of synthetic-aperture radar in measuring snowpack water equivalent. Side-looking airborne radar (SLAR) has been used for observing ice cover over the Great Lakes and ocean areas. Such systems can provide high resolution imagery (tens of meters) over sizeable swath widths (tens of kilometers) with no scale variation across the image.

IV.3.5 Satellite Observations of Snowcover

Studies of the application of satellite imagery to snowcover analysis began in the 1960's. A brief state-of-the-art review is provided herein. However, rapid changes in technology will soon render this summary obsolete. Continued sensor development will undoubtedly result in new techniques applicable to snowcover analyses. For example, at the present time a microwave scatterometer is currently being tested in sea-ice and sea-state experiments and eventually the sensor might be applicable to the analysis of the physical properties of a snowcover.

In 1977 the two principal satellite systems used for snowcover studies and operational analysis in North America were the polar orbiting Landsat and NOAA systems. The geostationary SMS/GOES satellites can provide useful data for snowcover analysis south of about 50°N, with an excellent time resolution of 30 minutes, but the viewing angle causes distortion problems for Canadian applications. The resolution of a few kilometers over southern Canada is not as good as that provided by Landsat or NOAA.

The first Earth Resources Technology Satellite (ERTS) was launched in 1972 and later renamed Landsat 1. Landsat covers a surface swath width of 185 km with a resolution of about 90 m at nadir. The Multispectral Scanner Subsystem (MSS) provides data in four spectral bands:

- MSS 4: 0.5 - 0.6 μm - green
- MSS 5: 0.6 - 0.7 μm - red (strong reflectance of dry and melting snow; similar to NOAA-VHRR visible band)
- MSS 6: 0.7 - 0.8 μm - near infrared
- MSS 7: 0.8 - 1.1 μm - near infrared (least water penetration, low reflectance from melting or metamorphose snow)
This multi-spectral imagery can be used separately, to obtain representative signatures in each of the four bands, or combined to give black and white or false colour composites. Landsat provides data for a given area at mid-latitudes once every 18 days. Convergence of swaths at high latitudes provides for more frequent coverage. When two Landsats are in orbit a temporal resolution of twice every 18 days is provided, but even coverage at nine day intervals may be insufficient for many applications in operational hydrology over southern Canada and, in particular the Prairies. A significant proportion of the imagery (typically 50% or more) is unusable because of cloud cover. In addition, the data were not available in real time in 1977. "Quick-look" imagery could be obtained within about ten days, while corrected and annotated images were available within two or three months. In general, Landsat data are being used for research and to supplement NOAA VHRR and conventional ground data for operational snowcover analyses.

Landsat also has a data retransmission capability. It can collect surface observations for example, from precipitation gauges or snow pillows, and from automatic telemetering stations, and retransmit them to distant data collection centers. This capability is useful in areas of significant snow storage that are remote or relatively inaccessible where it may not be feasible to install manned observing stations. With the Landsat system, simultaneous remotely-sensed imagery and ground based information can be collected, eliminating the significant problem of time lag in correlating the observations.

The NOAA satellites provide twice-daily coverage (mid-morning and late evening over southern Canada) of all areas in a swath width of about 1900 km, a width where distortion due to the earth's curvature is acceptable for snowcover analysis. The NOAA-5 satellite in orbit in 1977 carried two observing systems, the Scanning Radiometer (SR) and the Very High Resolution Radiometer (VHRR). Both sensors provide data in the visible (0.6 - 0.7 μm) and the thermal infrared (10.5 - 12.5 μm). The resolution of the VHRR is 900 m at nadir, while the SR has a maximum resolution of 3 km.

IV.3.5.1. Recent snow studies and current applications. In 1968 the World Meteorological Organization initiated a project on "snow studies by satellites", and subsequently published a status report and a report on an International Seminar (WMO, 1973, 1976) which incorporated results from a number of countries. The main emphasis in this project was the determination of the areal extent of snowcover, expressed as a fraction of the total basin area that was snow-covered to a depth of at least 5 cm. In a few cases the feasibility of determining other parameters, such as snow depth, was also examined. Canadian studies in support of the WMO project were conducted on the Saint John, Souris, Lake-of-the-Woods and Columbia basins. In general these studies were based primarily on NOAA-VHRR data (Ferguson and Lapczak, 1977) using limited "ground truth" information. Landsat imagery was used to supplement the analysis and image analyzers (density slicers) were employed to carry out grey scale analyses of the images. Optical equipment was used to superimpose original or density sliced images on base maps. The feasibility of determining areal extent of snowcover by the method was demonstrated. Its accuracy is highest over non-forested terrain and lowest over areas of dense coniferous forests where the snowcover may only be visible in larger clearings.

In some situations correlations have been found between image brightness and snow depth (McGinnis et al, 1975, Ferguson and Lapczak, 1977) but interpretation problems are formidable and a general operational technique has not been developed. Because of the inherent limitation of the penetration of visible light into the snowcover it seems likely that the determination of snow depth by remote sensing will have to rely on other portions of the spectrum, particularly the microwave wavelength band.

Operational snowcover mapping is being carried out by U.S. National Environment Satellite Service. A precision of ± 5% is claimed for areas greater than 5000 km² (Wiesnet, 1974).

McGinnis et al (1975) suggest the improved VHRR of later NOAA satellites permits mapping of snow limits to better than 10 km.

It has been recommended (WMO, 1976) that for operational hydrological purposes Landsat should be applied to basins with areas of 10 km² or larger, NOAA VHRR imagery should not be used on watersheds smaller than 1000 km² and the minimum basin size for NOAA SR data.
should be of the order of 10,000 km². For hydrologic applications it would be desirable to adapt forecasting models to accept remote sensing inputs.

Satellite snowcover applications also include regional or hemispheric delineations of snow and ice cover. Hemispheric maps of snowcover are distributed weekly by the National Oceanic and Atmospheric Administration (NOAA, 1977) which also prepares 10-day composite minimum brightness charts for detecting hemispheric snowcover under changeable cloud conditions. These charts are updated daily, recorded on tape and displayed as photographic images.

At present, relatively simple optical-electrical interpretation techniques are being used for operational snowcover mapping using satellite data. These methods involve elements of operator skill and subjectivity. More sophisticated and objective digital analysis methods have the potential to produce more accurate results, but they are still in the research and development stage. Suitable cost-effective digital analysis routines (based on the raw data with appropriate correction factors) will likely develop over the next few years. In that event, the imagery will still be useful for visual comparison. Ground based information will still be needed for calibration and verification and for filling gaps in the satellite data. Satellites are thus seen as one component of a total observing system which will continue to include surface observations.

V. SUMMARY

This paper attempts to document the "state-of-the-art" on evaluating areal snowcover accumulation and distribution. A review of the major processes and factors which govern accumulation patterns under mountain-forest and prairie environments is presented. Consideration is also given to methods of simulating and measuring the areal differentiation of snowcover.

From this review it is evident that at the present time, because of the lack of knowledge of the snow transport and deposition processes and the complex nature of the accumulation phenomenon, it is impossible for the hydrologist to define by a physically-based, mathematical model snowcover distribution patterns. Hence, certainly for the immediate future operational hydrologists will continue to rely very heavily on the use of ground measurements combined with empirical relationships and other measurement techniques for obtaining estimates of basin snowcover. There is an urgent need however, for hydrologists and other users of snowcover data to establish the sensitivity of the system output to space-time variable snowcover inputs. For example, for streamflow forecasting purposes, what level of precision is needed in defining areal snowcover distribution to optimize streamflow prediction? Such information would serve the useful purpose in defining the sophistry and effort that need to be given to the snow accumulation and distribution problem.

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