PRAIRIE SNOWMELT RUNOFF

D.M. Gray[1], J.W. Pomeroy[2] and R.J. Granger[3]

ABSTRACT

The research needs required for the development of an operational snowmelt model for the Prairie environment are presented. Major emphasis is placed on the processes of snowcover accumulation, snowmelt and meltwater infiltration.

The current status of models under development by the Division of Hydrology, University of Saskatchewan for describing snow transport and the blowing snow phenomenon, estimating the net radiative flux to a snow surface, calculating snowmelt quantities using an energy balance approach, and evaluating meltwater infiltration to frozen soils are briefly described. Future studies required to refine and verify these systems so they may be applied in practice are discussed.

---

[1] Chairman and Professor, [2] Graduate Student and [3] Research Officer, Division of Hydrology, University of Saskatchewan, Saskatoon, Saskatchewan, S7N OWO

INTRODUCTION

This paper discusses the current "State-of-the-Art" and the research needs for the development of an operational snowmelt runoff model for the Prairie environment. Special emphasis is given in the discussions to the processes of snowcover accumulation, snowmelt and meltwater infiltration to frozen ground.

SNOWCOVER ACCUMULATION

Reliable information on the amount and distribution of snow water over an area at the time of melt is central to improved forecasts of the volumes and rates of runoff from snowmelt. Such data establish the potential runoff volume and have a direct bearing on the spatial depletion of snow-covered areas, the rate of melt, the peak flow and the area of a watershed contributing to flow. Prairie hydrologists are well-aware of problems in streamflow forecasting caused by the variable nature of the "Contributing Area". Large areas of a watershed may not contribute to flow because of the lack of snowcover and large volumes of depressional storage. However, even in low snow years significant runoff can occur, the source being snow in the less visible channels and depressions that feed the main drainage-way. Generally, the size of the "Contributing" area tends to increase with an increase in the depth of snowcover and antecedent moisture conditions. Further research is needed into the interactions of snowfall amounts and distribution, soil infiltration potential, drainage-basin characteristics and contributing area.

The accumulation of a prairie snowcover is not a straight-forward process, as fallen snow undergoes redistribution by the wind. Steppuhn and Dyck (1974) have shown however that snow distribution patterns are similar over surfaces of similar aerodynamic characteristics. For purposes of predicting snowcover accumulation on various surfaces for a prairie snow-melt model, it is felt that modelling the aerodynamics of snow transport is required. Because of the complexity of the phenomenon, such a model must necessarily be a computer simulation of erosional and depositional processes over a winter. The Division of Hydrology, University of Saskatchewan is attempting to describe the physics of drifting snow and integrate this understanding into a model of snow transport. While this is not the only viable approach, the lack of a suitable database on the Canadian Prairies for calculating snow transport limits the accuracy of empirical models. Research needs pertinent to specific components of a physically-based model are addressed below.

The snow accumulation model presently being developed calculates snow transport fluxes using data collected hourly by Atmospheric Environment Service stations and a geographical information grid of land use, vegetation and physiography. Snow transport is calculated for fully-developed flow conditions which exist in the centre and down-wind portion of an aerodynamically-uniform terrain. The process is modelled using a control volume of unit horizontal area 10 m in height (see Fig. 1) and a mass balance used to determine the erosion/deposition rate. That is,

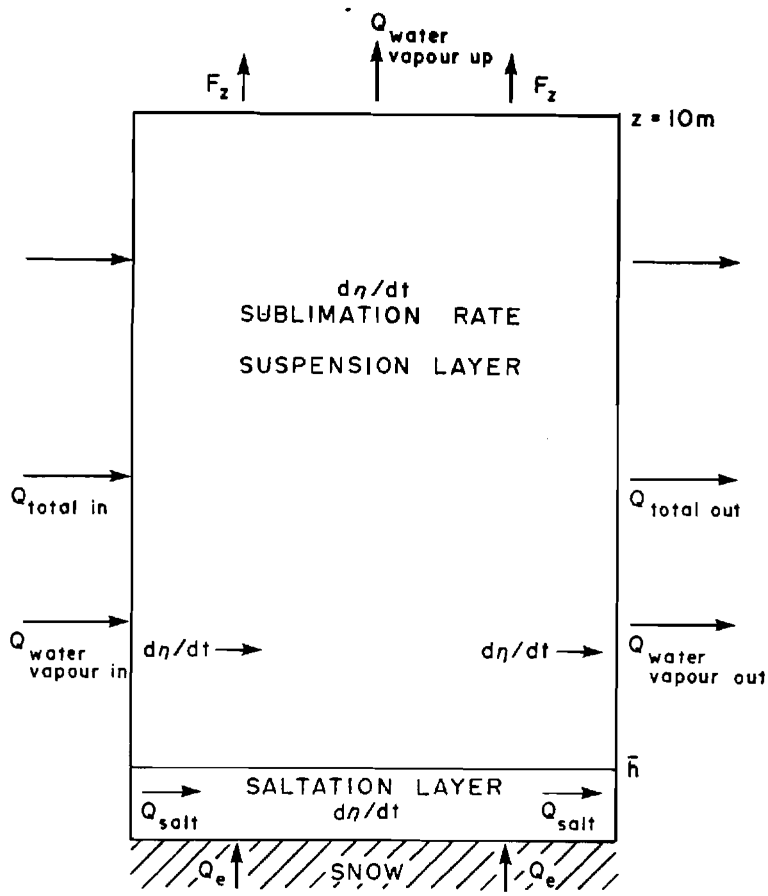

Figure 1. Control volume illustrating the snow and water vapor fluxes in fully developed blowing snow.

[1] $$Q_e = Q_{out} - Q_{in} + d\eta/dt + F_z,$$

where Q_e is the erosion rate (if negative the deposition rate), Q is the horizontal mass flux of drifting snow in and out of the control volume, $d\eta/dt$ is the sublimation rate of drifting snow within the volume and F_z is the vertical flux of snow at 10 m. For fully-developed transport the horizontal snow flux entering the control volume equals the flux leaving. The snow surface erosion/deposition rate is therefore equal to the change of snow mass in the volume due to either vertical transport or sublimation per unit time.

Two-dimensional, Fully-developed Flow

A model describing two-dimensional, fully-developed flow allows evaluation of the erosion rate in the centre of an aerodynamically-uniform terrain.

Horizontal transport of snow by the wind is analogous to aeolian and fluvial sediment transport and equations for drifting below 2 cm height (saltation) based on Bagnold's (1973) work have been developed by Pomeroy and Male (1985; 1986) and Schmidt (1982; 1986). The mass flux of saltating snow passing through a one meter wide column perpendicular to the flow direction is directly related to the windspeed cubed and inversely related to the aerodynamic surface roughness and the threshold windspeed at which particles are first ejected from the snowpack. These two variables index the resistance of the terrain and snow surface to drifting. Research needs exist in parameterizing these factors in terms of readily measured quantities. For example, vegetation height, density and patterns all affect terrain resistance to drifting, but quantitative relationships between these factors and the aerodynamic surface roughness during blowing snow are few. The meteorological and snowpack metamorphic processes producing cohesion of crystals at the snow surface are not well understood. No predictive models for this cohesion have been developed.

Blowing snow above a few centimeters in height travels in suspension and theory describing this mode of transport has been developed by Budd (1966) and operationalized by Pomeroy and Male (1986). The mass flux of suspended snow passing through a one-meter wide column perpendicular to the flow direction is a function of the windspeed, degree of atmospheric turbulence and the mass density of saltating snow in the air (drift density). The suspended mass flux is roughly 1/10th the saltating flux at low windspeeds (7 m/s) and increases to over 1/2 the saltating flux at higher windspeeds (18 m/s).

The sum of saltating and suspended mass fluxes equals the horizontal flux of blowing snow which has a distribution that varies with height, windspeed and surface conditions as shown in Fig. 2. It is evident in those data that the horizontal flux drops off rapidly with height, suggesting most transport is below 1 m (10 m for extreme conditions). The effects of differing threshold conditions and windspeed on the total horizontal snow flux are shown in Fig. 3 in which the flux is plotted against the 10-m wind speed, u_{10}. Differing threshold windspeeds for snow drifting result in a wide variation of horizontal fluxes at a given windspeed, especially when these fluxes are low; conversely, the horizontal fluxes are relatively uniform when the fluxes are high. Input parameters for the calculation of the horizontal mass flux are the threshold windspeed for drifting, ambient windspeed and the aerodynamic surface roughness. Wind speed is routinely measured at most climatological stations and the surface roughness can be estimated from the geometry of exposed vegetation using techniques of Lyles and Allison (1976) and Lettau (1969). The results from field experiments conducted near Loreburn, Saskatchewan (Pomeroy and Male, 1986) have allowed preliminary confirmation of the horizontal mass flux calculation for fully-developed flow over a snow-covered, summerfallow field.

Research needs regarding the determination of the fully-developed horizontal mass flux of wind transported snow are: further measurements of the saltation flux over exposed stubble and other vegetation; the development of a shear-force partitioning theory to describe this flow; and the development of techniques for predicting the threshold windspeed for a snow surface based on physical properties of the snowpack and for relating these snowpack properties to land use and antecedent weather conditions.

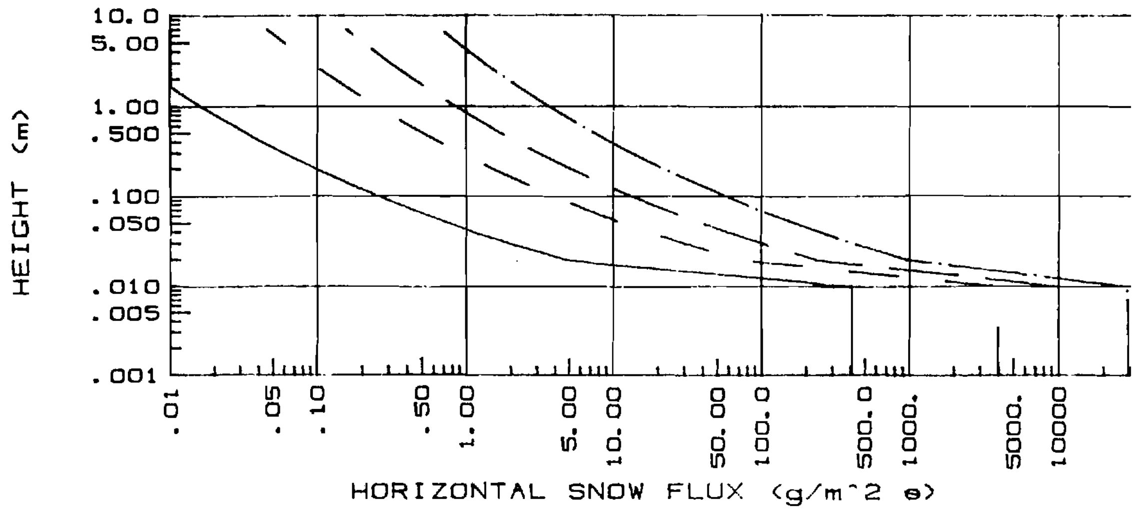

Figure 2. Vertical profiles of the horizontal mass flux of blowing snow. ——— u_{10} = 5 m/s, ---- u_{10} = 7.5 m/s, —— —— —— u_{10} = 10 m/s, —— • —— u_{10} = 15 m/s.

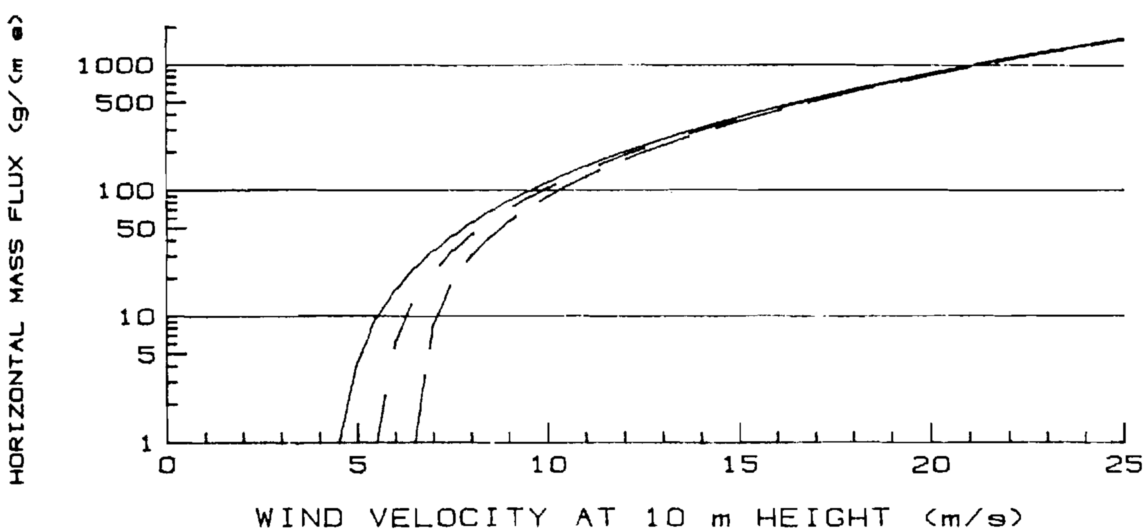

Figure 3. Horizontal mass flux of blowing snow to a height of 10 m. ——— threshold u_{10} = 4.5 m/s, ---- threshold u_{10} = 5.5 m/s, —— —— —— threshold u_{10} = 6.5 m/s.

The vertical mass flux of suspended snow at some height, is a function of the drift density at that height, the saltating drift density and the degree of atmospheric turbulence (Pomeroy and Male, 1985). Atmospheric turbulence is determined by the windspeed and aerodynamic surface roughness. The applicability of current calculations to the case of blowing snow has not been verified and field measurements of windspeed, drift density and particle size profiles with height are required for confirmation, or alternatively to suggest a more appropriate approach to the problem.

Sublimation of blowing snow can be a significant hydrologic loss, exceeding several mm of water equivalent per day in some events. In the model, the sublimation rate is calculated using an involved heat and mass transfer equation developed by Schmidt (1972) and modified by Lee (1975). The rates of sublimation depend on the drift density, atmospheric temperature and humidity. For example the sublimation rate from a quantity of drifting snow increases 25 fold when the temperature increases from -35° to -1° C. If vertically-transported snow (above 10 m) is considered to eventually sublimate, the sublimation rate equals the erosion rate in fully-developed flow and can amount to over 30 mm of snow water equivalent per day in a severe storm. Measurements of vertical profiles of temperature and humidity at Loreburn confirm substantial rates of water vapor production in the lower 0.5 m of the atmosphere during blowing snow, with relative humidities 20% higher at 120 mm than at 2 m. However, confirmation of the calculated sublimation rates by direct field measurements has not yet been made and is needed.

Two-dimensional Developing Flow

Developing flow exists where changes in windspeed, snow surface resistance or aerodynamic resistance have caused the flow through the control volume to be unrepresentative of both the flow at the top of and the resistance at the bottom of the volume. As a result the incoming and outgoing horizontal mass fluxes for a volume are unequal. This effect can result in erosion/deposition rates several orders of magnitude different than those expected for fully-developed flow. Observations in Saskatchewan indicate developing-flow zones as small as 2 m downwind from a land-use change from fallow to stubble and over 300 m downwind from a wooded lot. An experiment to be performed near Saskatoon in the winter of 1986-87 will measure the increase in the horizontal mass flux downwind from a shelterbelt. The results of this experiment should allow calculation of the erosion rate using unequal horizontal mass fluxes. There is also a need to measure the change in blowing snow fluxes as flow proceeds over uneven topography. These measurements must be explained in terms of models of internal boundary layer growth to provide a general model of snow-air flow over irregular terrain.

Summary

In summary, there is need for research on the processes of transport and sublimation of snow during transport. While there has been good progress in understanding and modelling snow transport over flat, homogeneous terrain, useful applications of these models will require algorithms to model snow transport near terrain boundaries and over irregular terrain. Measurements of meteorological variables and blowing snow fluxes should be directed toward evolving and calibrating aerodynamic models of flow in these conditions. For improved operational application of transport modelling, concepts such as equilibrium snow depths as a function of windspeed and snow surface hardness for a particular surface roughness should be used. A detailed geographic information network of land use, terrain and vegetation would improve the results of the snow accumulation model and could be used in calculating other components of snowmelt.

In time, one can perceive a model capable of describing snow accumulation over areas of mixed land use, vegetative cover and terrain. When combined with developing technologies for estimating snow water equivalent with satellite imagery one may look forward in the future to vastly improved methods for estimating snowcover and snow water distribution on a spatial scale.

SNOWMELT

Because snowmelt represents the phase change of ice to water, the physical framework for calculating this quantity is the energy equation which sums the radiative, convective, advective and conductive fluxes to and the rate of change in internal energy of a control volume of snowcover (Male and Gray, 1980). Improvements in predicting melt quantities are most likely only with the development of simple operational procedures that provide reliable estimates of the different terms affecting the energy available for melt. Major research efforts need to be directed to developing these methodologies.

As shown by Granger (1977) when advective fluxes are negligible, net radiation and the transfer of sensible heat govern the melt of a shallow Prairie snowcover. Generally it is found at the beginning of "active" melt of a complete snowcover that net radiation is the dominant flux for melt, whereas later in the sequence energy supplied from both sources may share equal importance. Only in the cases where the exchange is dominated by energy transferred from a large, stagnant, slow-moving warm air mass or derived over patchy snowcover conditions will simple procedures, such as the temperature index approach commonly applied to forested, mountainous regions, give reasonable estimates of melt. It is because the relative amounts of energy available for melt from radiative and convective sources vary widely from day to day and the fact that over open grassland areas net radiation and ambient air temperature generally show poor correlation that the temperature index approach can not be applied with confidence.

Over the years the possibility of using an energy balance approach for modelling snowmelt in operational models has been questioned. Some of the concerns have evolved from the assumed complexity of the approach, the lack of measurements and reporting stations, the difficulty in measuring different parameters and evaluating the different components and others. In the discussion below readily measured meteorological variables are employed in simple empirical relationships to demonstrate that estimates of daily net radiation of sufficient accuracy for operational models are possible.

Net Radiation

Net radiation is the sum of the net short-wave flux absorbed by the snowcover (Q_{sn}) and the net long-wave flux at the surface (Q_{ln}), i.e.,

[2] $$Q_N = Q_{sn} + Q_{ln}.$$

Davies (1965, 1967) and Davies and Idso (1979) have demonstrated from analyses of both daytime and nightime values of net radiation that over bare and vegetative surfaces Q_{sn} is the dominant flux and plots of Q_N and Q_s - the incident short-wave flux - are remarkably linear with high correlation coefficients. These findings suggest that variations in albedo of vegetative surfaces during summer are small and albedo does not vary widely under clear-sky conditions. Further, Davies and Buttimor (1969) noted that the linear association could also be applied to sky conditions when the incoming long-wave radiation is not constant.

Figure 4 shows a plot of measured values of Q_N and Q_{sn}, observed at the Bad Lake Climatological Station located in western Saskatchewan during periods of "active" melt of a Prairie snowcover and during the "Postmelt" period immediately following the disappearance of the seasonal snowcover for the period, 1972-1985, inclusive. Note, Q_N are daily values from a Funk (Middleton) net radiometer and the Q_{sn}-values were calculated as the difference between global incoming (Q_s) and reflected (Q_r) short-wave radiation obtained from upright and inverted Kipp solarimeters. The best-fit equation of the line is:

[3] $$Q_N = -0.371 + 0.522\ Q_{sn}$$

which has a correlation coefficient of 0.89 and a standard error of estimate of 1.39 MJ/m^2-d with Q_N and Q_{sn} are in MJ/m^2-d. Considering the high degree of association between the variables (despite the fact Q_{sn} was calculated as the difference between two point measurements), the sensitivity of a snowmelt runoff model to varying inputs of Q_N can not be established at this time, and that 1 MJ/m^2-d of energy applied to ice at 0°C will produce a depth of melt of approximately 3 mm, it is anticipated the estimates of Q_N given by Eq. 3 will satisfy operational requirements.

Incident Short-wave Radiation

To use Eq. 3 requires an estimate of Q_{sn}. Unfortunately measurements of Q_s and Q_r are not usually taken at most reporting climatological stations. Although the lack of data is a problem, reasonable estimates of the parameter for Prairie conditions may be possible using empirical procedures. Q_{sn} is given by the expression:

[4] $$Q_{sn} = Q_s(1-A)$$

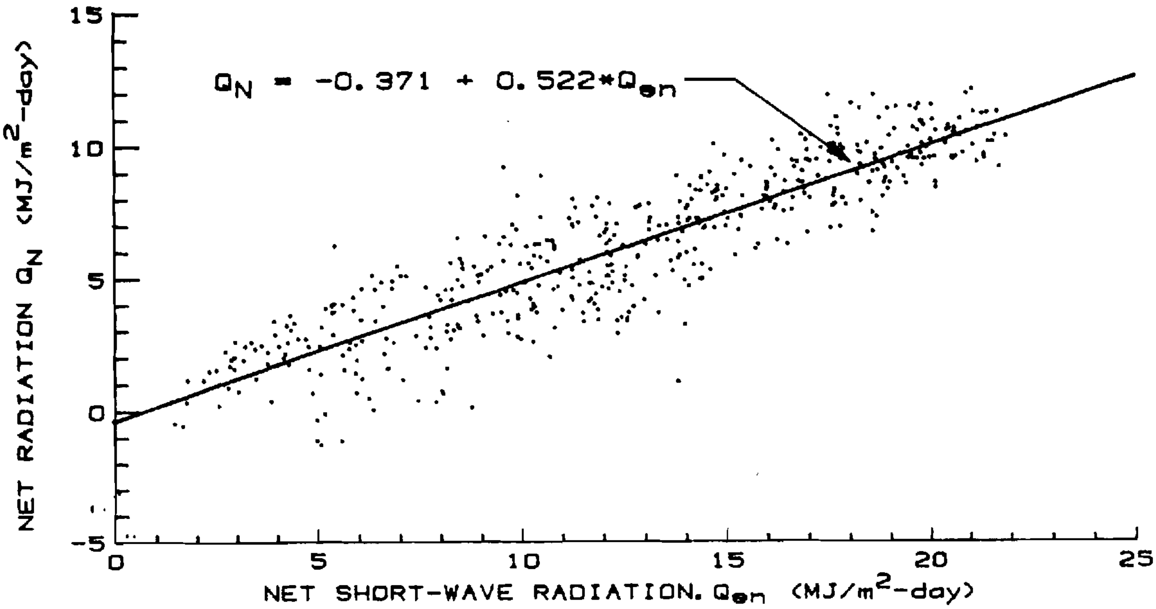

Figure 4. Relationship between daily net radiation and net short-wave radiation during Melt and Postmelt periods, 1972-1985 inclusive, Bad Lake, Saskatchewan,

in which Q_s is the incident short-wave flux to the snow surface and A is the albedo. Numerous investigators (Penman, 1948; Mateer, 1955; Brutsaert, 1982) have shown that Q_s can be related to either the clear-sky insolation (Q_o) or the extraterrestrial radiation flux (Q_A) and the sunshine ratio (n/N) by a simple linear equation of the form:

[5] $$Q_s = Q_o(a + b(n/N) \text{ or } Q_A(A + B(n/N)$$

in which the coefficients a, A, b and B must be evaluated from measured data and n/N is the ratio of the number of bright sunshine hours (for example, as measured by a sunshine recorder) to the number of possible hours of sunshine. The advantage of Eq. 5 is its simplicity; once the values for the coefficients have been assigned only a measurement of "n" is needed for a solution, as Q_o, Q_A and N are fixed in time and geographical location. Use of the clear-sky insolation as input to the equation is probably the most amenable for snowmelt calculations because models have been developed that allow partitioning Q_o to direct beam and diffuse components. This partitioning permits accounting for the effect of slope on the energy received. Two expressions that may be used for this purpose are those proposed by Garnier and Ohmura (1970) and List (1968). The relationship between Q_s and Q_o - calculated by these procedures - for Bad Lake for the months of Feb. - Mar. inclusive is plotted as a function of n/N in Figure 5. The relationship is described by the equation:

[6] $$Q_s/Q_o = 0.518 + 0.519n/N,$$

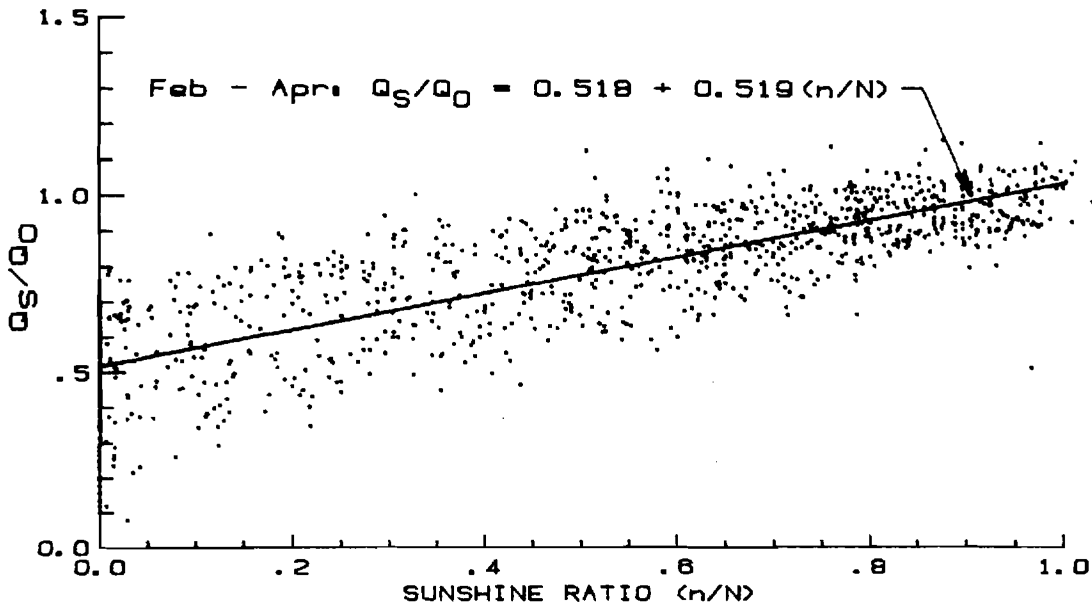

Figure 5. Relationship between ratio of measured daily global radiation to the calculated clear-sky, short-wave radiation and sunshine ratio for the months Feb.-Apr. inclusive at Bad Lake, Sask., 1972-1985.

which has a correlation coefficient of 0.81 and a standard error of estimate of 0.12. Note, Eq. 6 uses a transmissivity coefficient p = 0.85 for the earth's atmosphere to calculate Q_o; the magnitude of "p" will change with season and geographical location.

Albedo

To use the estimate of Q_s from Eq. 6 to calculate Q_{sn} by Eq. 4 requires knowledge of the albedo, A. As stated above this parameter is not routinely measured at most climatological stations and therefore is usually selected from tabulated material based on different physical characteristics of the snowcover such as colour, ripeness, wetness, age and other properties. The material below briefly describes a first attempt toward the development of an albedo model for the Prairies.

From examination of the change in albedo with time during late winter and early spring it was observed that when the effects of snowfall and melt-freeze events were neglected the decay tended to follow an orderly pattern in each of the "Premelt", "Melt" and "Postmelt" periods (see Fig. 6) (Gray et al., 1986). During Premelt, the months preceding the period of Melt when the seasonal snowcover depletes and disappears, the albedo decreases gradually with time at a relatively constant rate. Average annual daily rates of change in the period, assumed to start Feb. 1^{st} of any year, were measured in the range 0.004 - 0.009/d with an average of 0.0061/d.

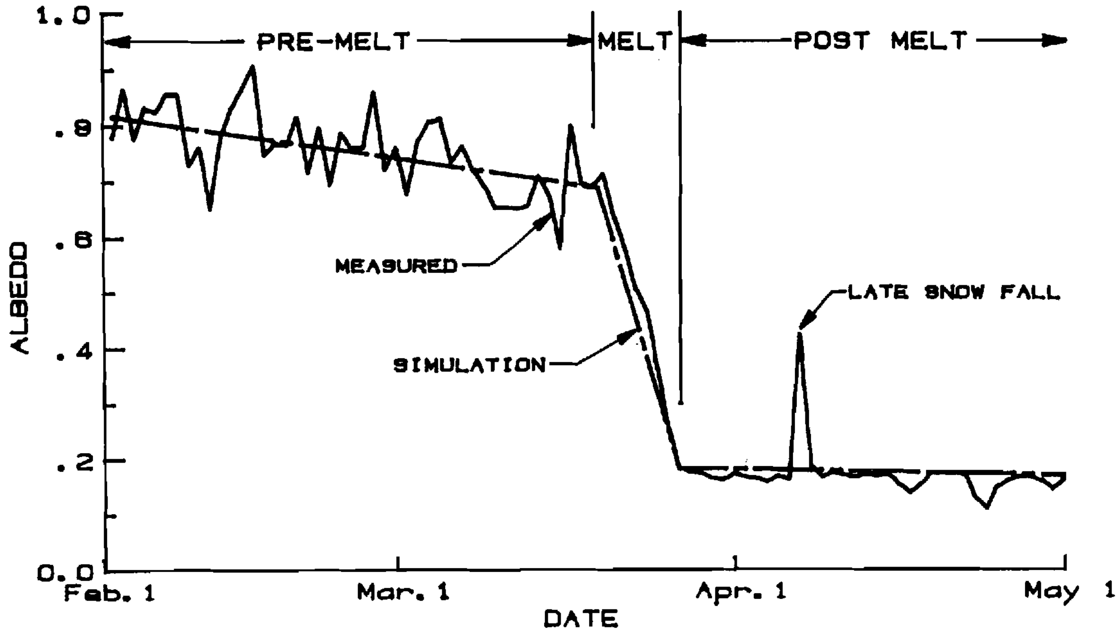

Figure 6. Schematic of the variation in albedo of a Prairie snowcover with time during Premelt, Melt and Post melt periods (after Gray et al., 1986).

During Melt the decrease in albedo is largest and most rapid and frequently occurs continuously and uninterrupted at a reasonably constant rate. Gray et al., 1986 have demonstrated with point and aerial measurements that during continuous ablation of a snowcover the rate of decay in A is of the order of 0.071/d and can be described by the linear equation:

[7] $$A = 0.66 - 0.071t$$

in which t is the number of days after the start of continuous depletion of the seasonal snowcover ($r = 0.81$). Usually a prairie snowcover disappears in 5-8 days.

During Postmelt, when the ground is bare, the albedo of the prairie terrain can be taken as 0.17.

The material above provides the basis of a "Prairie" albedo model which is currently under development and evaluation. Essentially it consists of 3 line segments having different slopes. An algorithm of the model has been written and modified to account for albedo changes during the transition periods, e.g. Premelt to Melt, and those caused by snowfall and discontinuous melt events. Also built into the algorithm are procedures for establishing the "start" of melt based on the ambient air temperature and calculated values of net radiation. Figures 7a and 7b taken from Gray and Landine (1986) compare the measured and calculated albedos for a high snow year - 1974; depth of snowcover at the start of melt ~46 cm and for a low snow year, 1980; initial snow depth ~9 cm. It can be noted that the

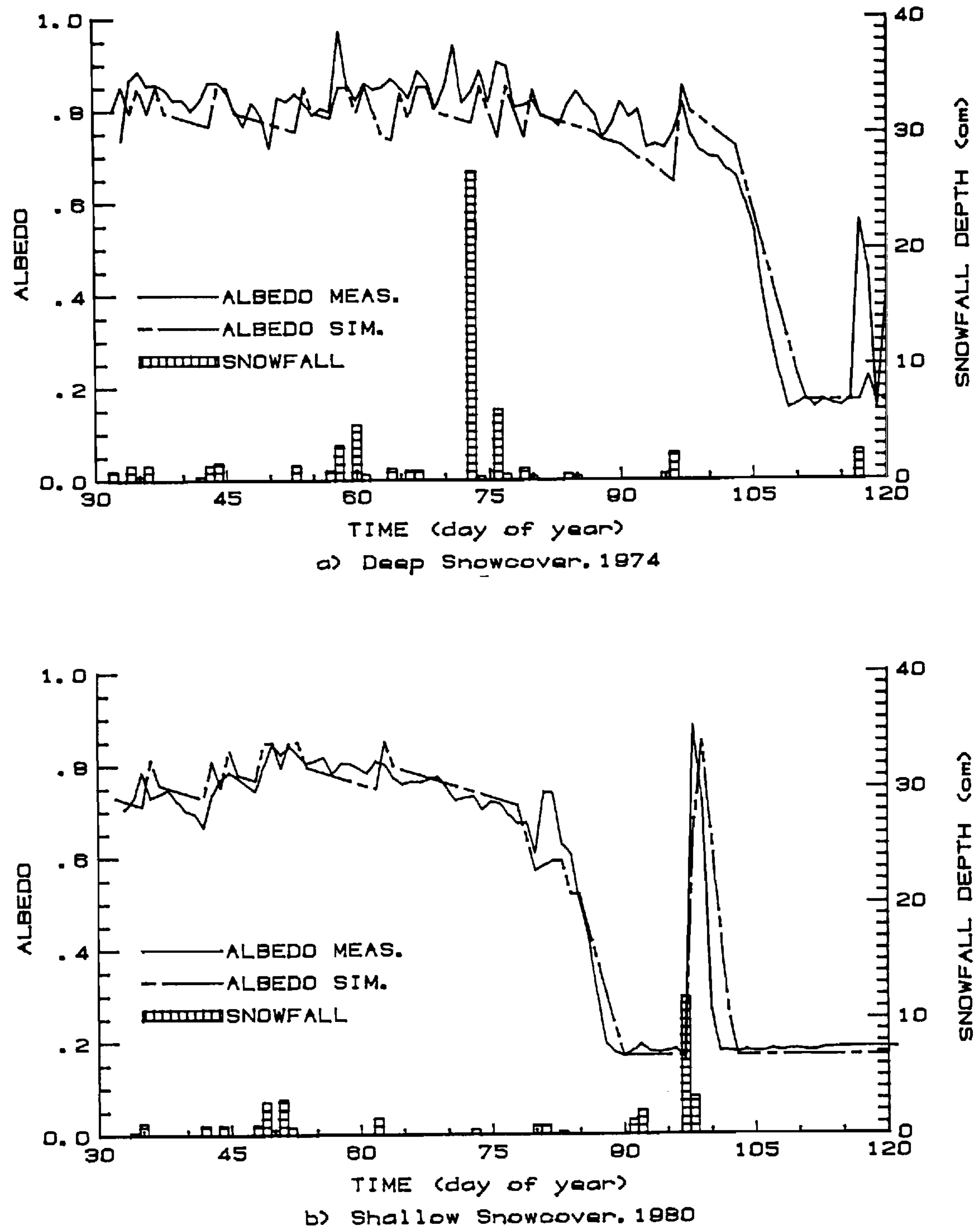

Figure 7. Comparison of measured and simulated albedo depletion curves for Bad Lake, Saskatchewan: (a) 1974, deep snowcover; (b) 1980, a shallow snowcover.

agreement between values is reasonable and it is expected that the calculated albedos would give acceptable estimates of Q_{sn}. Note, the albedo decay patterns in both years conform very closely to those assumed by the model and the high degree of association between the measured and calculated values of A is expected, provided the start of Melt is predicted with reasonable accuracy. Comparisons between model and measured values made for years when frequent, periodic melt events occurred throughout the winter and spring, which in some cases led to complete ablation of the snowcover, were not as good and point to further refinement of the model.

In summary, it is suggested that a simple model for calculating the daily net radiative flux required for snowmelt calculations is close to being perfected. The same can not be said about procedures that may be used to obtain the sensible heat flux. A priority research need is therefore the development of "bulk" methodologies for this purpose, especially for patchy, snowcover conditions.

SNOWMELT INFILTRATION INTO FROZEN SOIL

Meltwater generated by a snowcover is destined either to infiltrate the underlying soil, evaporate or to run off. Whether one's interest lies in the area of determining volumes or rates of runoff, an understanding of the infiltration phenomenon is essential because it is the major process governing the apportionment of the meltwater to runoff.

When one is dealing with infiltration of meltwater on the Prairies one is usually dealing with infiltration into frozen soil. This process involves the complex phenomenon of coupled heat and mass transfer through porous media and is affected by many factors including the hydrophysical and thermal properties of the soil, the soil moisture and temperature regimes, the rate of release of water from the snowcover, and the energy content of the infiltrating water. In the absence of major structural deformations e.g., cracks or other macropores, the major hydrophysical property of a frozen soil governing its ability to absorb and transmit water is its frozen moisture content. The existence of an inverse relationship between infiltration and the frozen soil mositure content has been demonstrated or postulated by many investigators.

Implicit in the consideration of thermal properties is the definition of what constitutes a frozen soil. For example, findings reported by several Soviet investigators suggest that soils frozen to depths less than 500 - 600 mm may have infiltration characteristics similar to those of unfrozen soils, and as such could be viewed as being only "partially-frozen". On the Canadian Prairies, with the possible exception of areas subjected to "chinooks", the soils are usually frozen to depths greater than 600 mm at the time of snowmelt, and can be treated as "fully-frozen".

Although much progress has been made in modelling the phenomenon, infiltration into frozen soils must still be estimated from empirical equations or simple, physically-based relationships in solving broadscale water management problems.

Recent studies by the Division of Hydrology (Granger et al., 1984; Gray et al., 1985a, 1986a) have led to the development of a semi-empirical model describing snowmelt infiltration to frozen soils. The model is based on a comprehensive field study of infiltration to frozen soils in the Dark Brown and Brown soil zones, and on the results of infiltration studies under similar climatic regions of the USSR reported in the literature. The model assumes that frozen soils may be grouped into three broad categories with regard to their infiltration potential, namely: Restricted, Limited and Unlimited.

Restricted - infiltration is impeded by an impermeable layer, such as an ice lens at the soil surface or within the soil near the surface. For practical purposes the amount of meltwater infiltration can be assumed to be negligible and most of the snowcover water goes to direct runoff and evaporation.

Limited - infiltration is governed primarily by the snowcover water equivalent and the frozen water content of a shallow layer of soil, 0-300 mm.

Unlimited - dry heavily-cracked clays, coarse dry sands or other soils which contain a large number of large, air-filled, non-capillary pores or macropores at the time of melt. Most or all of the snow water will infiltrate.

It was found for medium to fine-textured, uncracked frozen soils to which entry of meltwater is not impeded by ice layers (i.e.: the Limited case) that: (a) the average depth water penetrated during the melt period was 260 mm (standard deviation = 100 mm) and (b) the amount of infiltration was relatively independent of soil texture and land use and inversely related to the average moisture content of the soil layer, 0-300 mm at the time of melt. These findings led to a set of relationships between snowmelt infiltration (INF), snowcover water equivalent (SWE) and the premelt moisture content (θ_p), which for cases where SWE $\geq$ INF can be approximated by the equation:

$$[8] \qquad INF = 5(1-\theta_p)SWE^{0.584}$$

in which INF and SWE are in mm and θ_p is the degree of pore saturation mm^3/mm^3. The equation has a correlation coefficient of 0.85 and a standard error of estimate of 5.5 mm. Figure 8 (taken from Gray et. al., 1986a) is a graphical presentation of the model.

Such a simple model of infiltration to frozen soil can be applied to a broad spectrum of water-related management problems where snowmelt and frozen soils are involved. To agriculturalists the model provides a useful tool for the planning of sensible snow management practices (trapping wind-transported snow) for augmenting soil water reserves needed for crop production in dryland farming areas (Gray and Granger, 1985; Gray et al., 1985b; Granger and Gray, 1986). The model has also been successfully applied to the problem of simulating streamflow from snowmelt (Gray et al., 1985a, 1985b). When interfaced with the U.S. National Weather Service River Forecasting System (NWSRFS) and the U.S. Army Corp of Engineers'

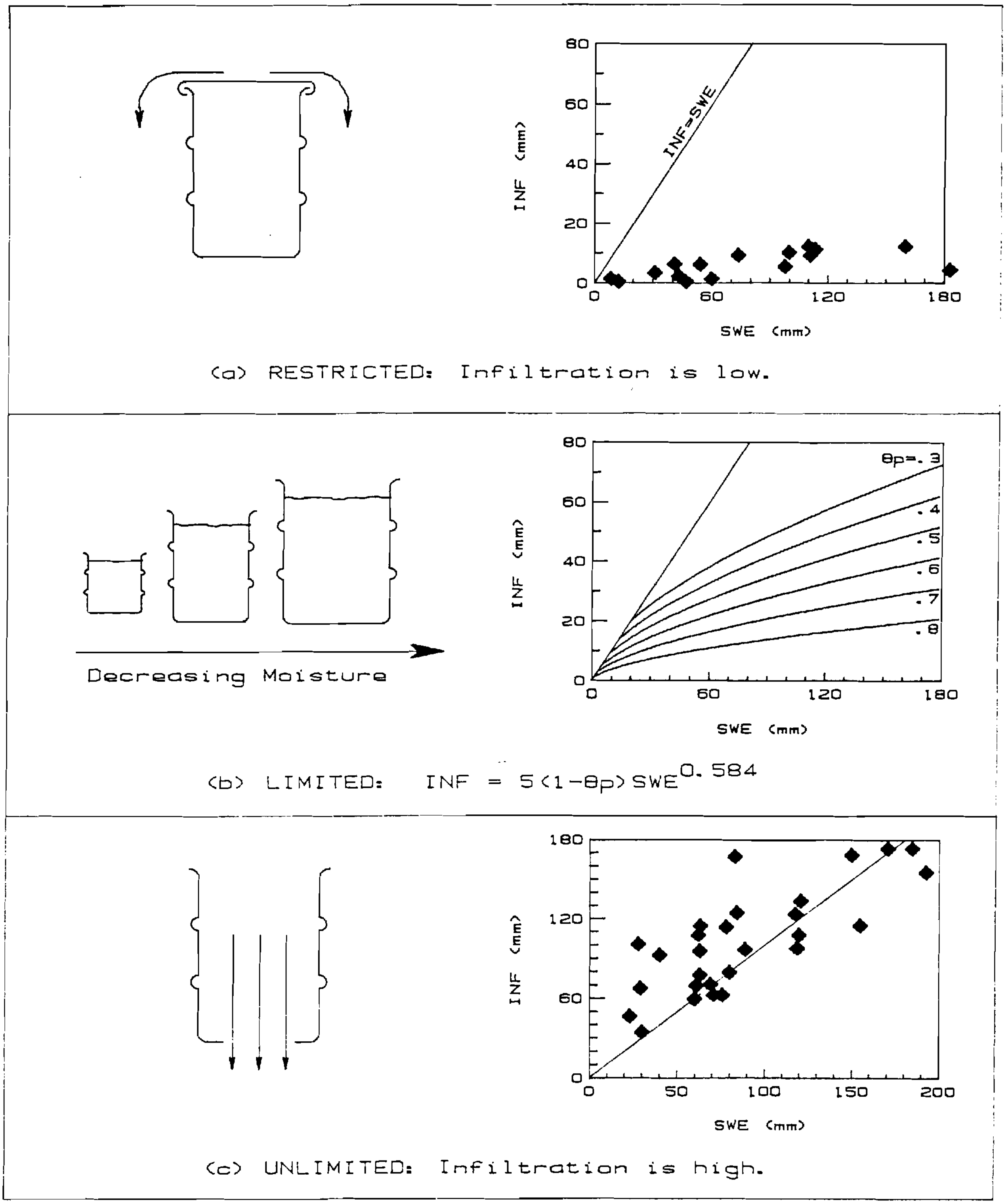

Figure 8. Snowmelt infiltration model for frozen Prairie soils: (a) Restricted, (b) Limited and (c) Unlimited (after Gray et al., 1986a).

Streamflow Synthesis and Reservoir Regulation System (SSARR) and applied to synthesizing flow on a Prairie watershed much closer agreement was obtained between the observed and simulated hydrographs with the infiltration model than with the original system. At its present stage of development the model must still be considered as "first-generation". However, because it is simple, has a physical base and does not require a large data base, it should in the interim find favor as a useful tool for water managers dealing with snowmelt and frozen soils.

There are several immediate research needs which would have a direct impact on the value of the infiltration model in operational practice. For example, information is needed on the range in moisture content at which different soils crack and the effects of late rains in the fall or melt events during the winter in changing the "infiltration potential" of a frozen soil. Additional verification of the algorithm is also needed. Further efforts should also be directed to study the infiltration process in "partially-frozen" ground where the depth of frost is limited, in areas subject to frequent mid-winter melts and in northern areas of discontinuous and continuous permafrost.

SUMMARY

The paper discusses the current State-of-the-Art and the research needs for the development of an operational snowmelt runoff model for the Prairie environment. Attention is given to the processes of snowcover accumulation, snowmelt and meltwater infiltration to frozen ground.

A snow accumulation model that calculates snow transport fluxes under fully-developed flow conditions from standard climatological data is described. Research is needed to quantify the saltation flux over different vegetation, to develop a shear-partitioning theory to describe the flow, and to derive techniques for estimating the threshold velocity for a snow surface from the physical properties of the snowcover. Further information is also required on changes to blowing snow fluxes caused by changes in land use and vegetation and variations in topography.

The fact that sublimation of blowing snow can be a significant hydrologic loss is emphasized and additional research on the process encouraged.

An empirical relationship that can be used to calculate the daily net radiative flux during melt and postmelt periods from the daily net global flux is presented. It is demonstrated that reasonable estimates of the incident short-wave flux can be obtained from existing models which make use of the solar constant, atmospheric transmissivity, actual sunshine hours and possible number of sunshine hours. A first-generation albedo model that allows partitioning the incident short-wave to absorbed and reflected quantities is described.

The major obstacle to the development of an energy balance model for calculating melt quantities is the lack of reliable methods for evaluating the sensible heat flux. A priority research need is the development of "bulk methodologies" for calculating this term, especially for patchy, snow-cover conditions.

A snowmelt infiltration model for frozen Prairie soils is described. The model groups frozen soils into three broad classes according to their infiltration potential, namely: Unlimited - soils having a large number of macropores (e.g. cracked soils); Limited - soils in which infiltration is governed by the snowcover water equivalent and ice/water content of the soil layer, 0-300 mm; and Restricted - soils in which an ice lens has formed on the surface or within the soil near the surface. References illustrating the application of the model in studies concerned with streamflow forecasting and meltwater enhancement to frozen soils are provided. The research needs specific to the model include information on the range in moisture content that different textured soils crack and the effects of fall rains and mid-winter melt events in changing the "infiltration potential" of a soil.

Further efforts in the general area of soil infiltration should be directed to studying the process in "partially-frozen" ground where the depth of frost is limited, in areas subject to frequent mid-winter melts and in northern areas of discontinuous and continuous permafrost.

REFERENCES

BAGNOLD, R.A. 1973. The nature of saltation and of "bed-load" transport in water. Proceedings of the Royal Society of London A, 332:472-504.

BUDD, W.F. 1966. The drifting of non-uniform snow particles. Studies in Antarctic Meteorology, American Geophysical Union Antarctic Research Series, 9:59-70.

BRUTSAERT, W. 1982. Evaporation into the Atmosphere. D. Reidel Publishing Company, London, England, 229 p.

DAVIES, J.A. 1965. The use of Gunn-Bellani distillator to determine net radiative flux for West Africa. Journal of Applied Meteorology 4:547-549.

DAVIES, J.A. 1967. A note on the relationship between net radiation and solar radiation. Quarterly Journal of the Royal Meteorological Society 93:109-115.

DAVIES, J.A. and D.H. BUTTIMOR. 1969. Reflection coefficients and net radiation at Simcoe, Southern Ontario. Agricultural Meteorology 6:373-386.

DAVIES, J.A. and S.B. IDSO. 1979. Estimating the surface radiation balance and its components. In Modification of the Aerial Environments of Crops (B.J. Barfield and J.F. Gerber, eds.). American Society of Agricultural Engineers, St. Joseph, MO. pp. 183-210.

GARNIER, B.J. and A. OHMURA. 1970. The evaluation of surface variations in solar radiation income. Solar Energy 13:21-24.

GRANGER,R.J. 1977. Energy exchange during melt of a prairie snowcover. M.Sc. Thesis, Department of Mechanical Engineering, University of Saskatchewan. 122 p.

GRANGER, R.J., D.M. GRAY and G.E. DYCK. 1984. Snowmelt infiltration to frozen prairie soils. Canadian Journal of Earth Sciences 21(6)996-677.

GRANGER, R.J. and D.M. GRAY. 1986. Combined subsoiling and snow management for drought attenuation. In Proceedings of the Canadian Hydrology Symposium No. 16-1986, Drought: The Impending Crisis, National Research Council of Canada, Associate Committee on Hydrology, pp. 213-225.

GRAY, D.M. and P.G. LANDINE. 1986. An albedo model for a Prairie snowcover. Internal Report, Division of Hydrology, University of Saskatchewan, Saskatoon, SK.

GRAY, D.M. and R.J. GRANGER. 1985. Snow management practices for increasing soil water reserves in frozen Prairie soils. In Watershed Management in the Eighties, Proceedings of the American Society of Civil Engineers, Denver, CO, pp. 256-263.

GRAY, D.M., J.W. POMEROY and P.G. LANDINE. 1986. Development and performance evaluation of energy balance snowmelt models. Final Report for Research Management Division, Alberta Environment, Edmonton, AB, Research Agreement RMD 83-34B, 67 p.

GRAY, D.M., P.G. LANDINE and R.J. GRANGER. 1985a. Simulating infiltration into frozen prairie soils in streamflow models. Canadian Journal of Earth Sciences, 22(30)464-472.

GRAY, D.M., P.G. LANDINE and G.A. McKAY. 1985b. Forecasting streamflow runoff from snowmelt in a Prairie environment. In Proceedings of the 7th Canadian Hydrotechnical Conference, Canadian Society of Civil Engineers, 1A:213-231.

GRAY, D.M., R.J. GRANGER and P.G. LANDINE. 1986a. Modelling snowmelt infiltration and runoff in a Prairie environment. In Proceedings of the Symposium: Cold Regions Hydrology, American Water Resources Association, pp. 427-438.

GRAY, D.M., R.J. GRANGER and W. NICHOLAICHUK. 1986b. Snowmelt infiltration to uncracked, cracked and subsoiled frozen soils. In Proceedings of the Snow Management for Agriculture Workshop, Swift Current, SK., July, 1985. (in press).

LEE, L.W. 1975. Sublimation of Snow in Turbulent Atmosphere. Unpublished Ph.D. Thesis, University of Wyoming, Laramie, WY. 162 p.

LETTAU, H. 1969. Note on aerodynamic roughness-parameter estimation on the basis of roughness-element description. Journal of Applied Meteorology 8:828-832.

LIST, R.J. 1968. Smithsonian Meteorological Tables. 6th Edition.

LYLES, L. and B.E. ALLISON. 1976. Wind erosion: the protective role of simulated standing stubble. Transactions American Society of Agricultural Engineers. 19:61-64.

MALE, D.H. and D.M. GRAY. 1980. Snowcover ablation and runoff. In Handbook of Snow: Principles, Processes, Management and Use (D.M. Gray and D.H. Male, eds.) Pergamon Press, Canada Limited, Willowdale, ON. pp. 360-436.

MATEER, C.L. 1955. A preliminary estimate of average insolation in Canada. Canadian Journal of Agricultural Science 35:579-594.

PENMAN, H.L. 1948. Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society A193;120-145.

POMEROY, J.W. and D.H. MALE. 1985. Physical modelling of blowing snow for agricultural production. In Proceedings of the Snow Management for Agriculture Workshop. Swift Current, SK. July, 1985. (in press)

POMEROY, J.W. and D.H. MALE. 1986. Wind transport of seasonal snowcovers. Chemical Dynamics of Seasonal Snowcover, NATO ASI Series, D. Reidel Publishing Co., Dordrecht, Netherlands. (in press)

SCHMIDT, R.A. 1972. Sublimation of wind-transported snow - A model. U.S. Forest Service Rocky Mountain Forest and Range Experiment Station Research Report, RM-90, Fort Collins, CO. 24 p.

SCHMIDT, R.A. 1982. Properties of blowing snow. Review of Geophysics and Space Physics, 20:39-44.

SCHMIDT, R.A. 1986. Transport rate of drifting snow and the mean windspeed profile. Boundary Layer Meteorology, 34:213-214.

STEPPUHN, H. and G.E. DYCK. 1974. Estimating true basin snowcover. In Advanced Concepts and Techniques in the Study of Snow and Ice Resources. National Academy of Sciences, Washington, DC. pp. 314-318.