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CLIMATOLOGICAL BULLETIN

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No. 13 April 1973

EDITORIAL FOREWORD

Recent developments in climatological research have been increasingly away from the traditional forms of large-scale analysis and the use of "normals" as a basis of generalisation. The subject has tended to move towards a treatment which aims at generalising from studies at time and space scales more in keeping with the actual climatic processes and influences found at the earth's surface. These developments are reflected mainly in two types of study: those which seek to evaluate climatic characteristics and influences on the basis of the frequency of occurrence of different categories of events occurring singly, or in combination, at appropriately short time intervals such as a day; and those which investigate, in detail, processes at the earth's surface in order to discover and model the important climatic factors and the time and space scales at which they operate. Examples of studies in the former category are those by Stephen Fogarasi concerning precipitation characteristics in the Arctic (see News and Comments in this Bulletin) and by Kala Swami on the role of daily and synoptic analyses in ecological studies (Bulletin No. 8, pp. 40-57).

The present number of the Bulletin offers four articles in the second category noted above. They are all concerned with hydrology, with particular reference to problems of snow and ice melting. As such they illustrate not only different aspects of climatological methodology but also the role realistic climatic studies can play in interdisciplinary research.

One of the main functions of the Bulletin is to draw attention to developments as they are in progress, and to encourage the presentation of studies and methods of research which are not yet at a stage suitable for offering to established journals but which are valuable indicators of current procedures. It is hoped that readers will find the four articles contained in this number valuable and that the articles will encourage others to offer material which they feel may be useful to the scientific community as indicators of work in progress.

B.J. Garnier

PHOTOGRAPHIC MEASUREMENT OF VEGETATION CANOPIES FOR USE IN THE COMPUTATION OF THE RADIATION BALANCE

by

John Fitzgibbon and T. Dunne*

For many purposes in hydrology, meteorology, and forest ecology, the radiation balance of the forest floor or of some other low-level surface under the canopy must be evaluated. This balance depends upon radiation conditions above the canopy and upon the characteristics of the canopy itself. Radiation in the open can be measured directly, or calculated from empirical relationships. The canopy characteristics vary with the age and species of vegetation, but also from place to place within an assemblage of similar vegetation. It is not possible to sample all the variations with direct measurements of radiation under the canopy. A simple way must therefore be found to measure the canopy characteristics in a fashion that is meaningful for radiation. These characteristics of the vegetation can be related to the measured radiation balance, and it is then quite simple to extend the calculation of the radiation balance by mapping the canopy parameters.

Past Developments

There have been a number of attempts to use photographic equipment to measure the effects of a vegetation canopy on the radiation balance. The amount of radiation reaching the forest floor is a function not only of the vegetation cover but also of the altitude and azimuth of the sun. Thus the technique of measurement must account for variation in the effect of the vegetation canopy over the whole hemisphere.

One commonly used instrument is the pin-hole camera, which produces a projection of the whole hemisphere and may be built quite cheaply. Another device used is a camera which photographs a concave or a convex mirror. They provide good results, and have been used to measure light

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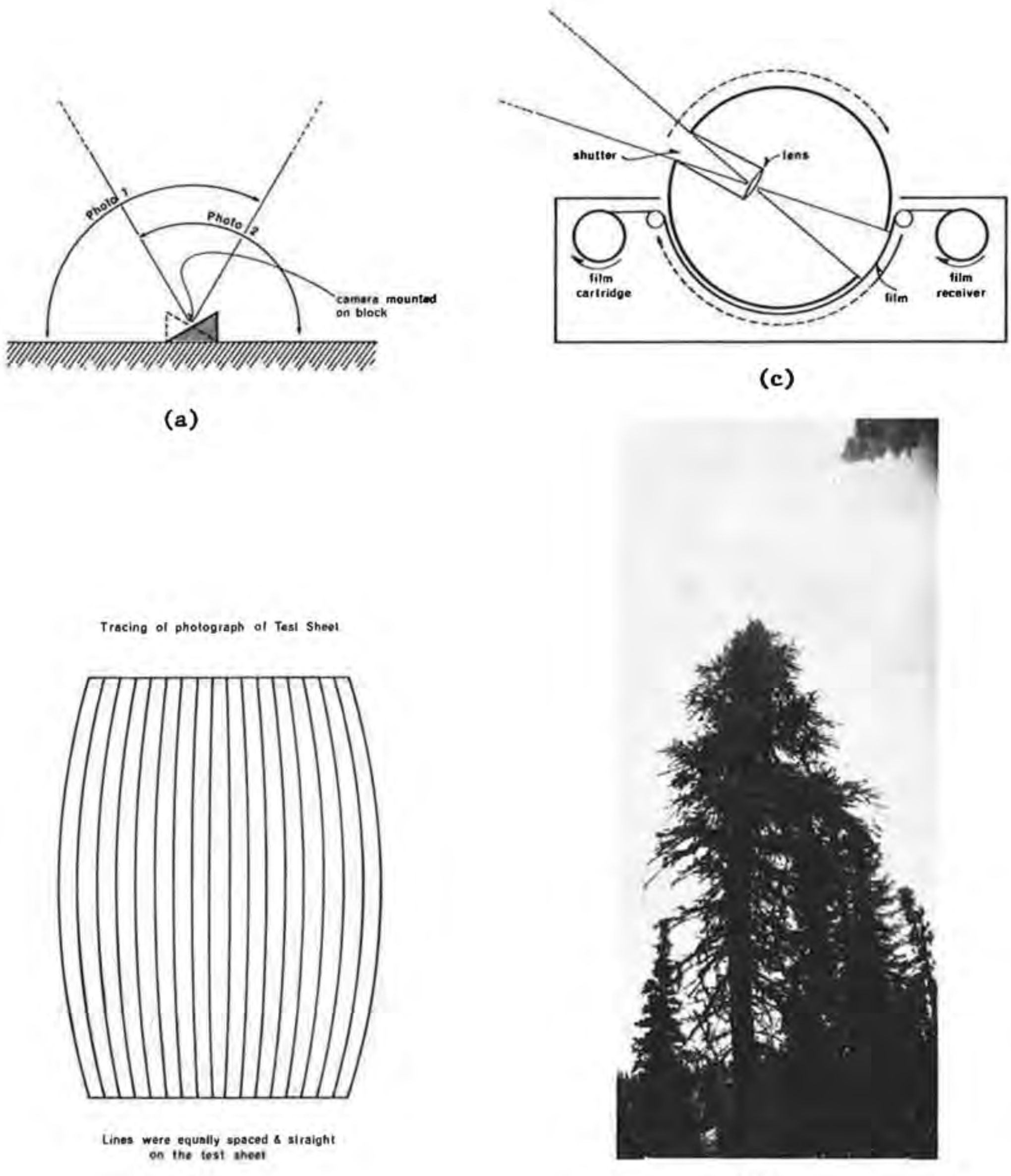


Fig. 1: Photographic Systems Used:
(a) photographic procedure;
(b) distortion of photograph;
(c) camera mounting;
(d) sample photo of open woodland.

in forests by Monsi and Saeki (1953). Such a camera, however, always appears in the photo, thus obscuring some of the area of the hemisphere. Anderson (1964) describes the use of the Hill camera for taking hemispherical photos of forest canopies of various types. The Hill camera produces a photo of the complete hemisphere as an equal area projection of the hemisphere which is analyzed by using a grid on the photo. Evans and Coombe (1959) have also used the Hill camera for canopy photography. Johnson and Vogel (1968) used a "fish-eye" lens to photograph forest canopies for various purposes. Such a lens only gives a portion of the hemisphere in the view and was used to develop an illumination index.

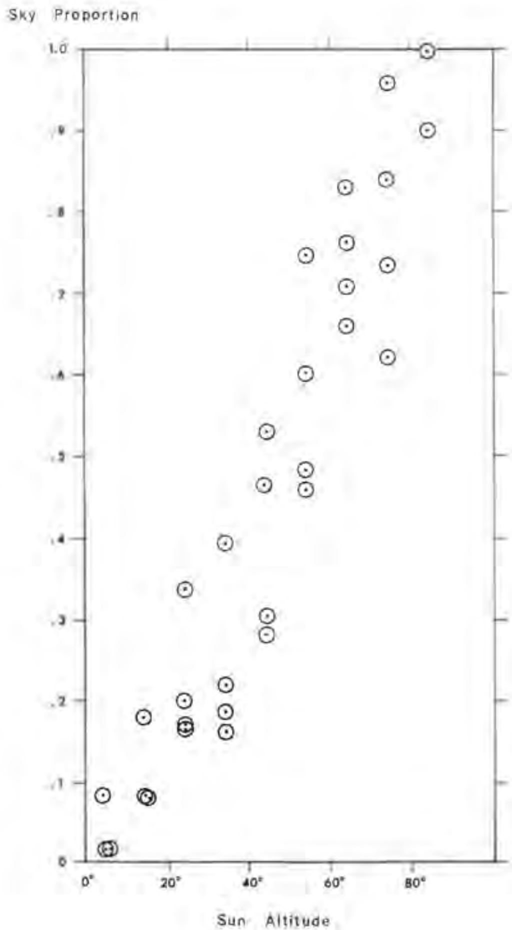
A New Method

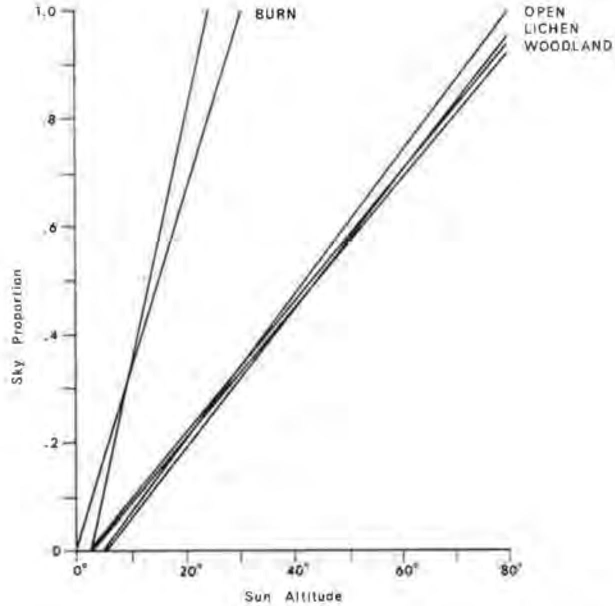
Two major approaches have been taken to the problem of measuring vegetation for the purpose of evaluating the effect of the canopy on the radiation balance. These are measurement by photograph, and measurement of silvicultural parameters of the vegetation (tree height, tree density, breast height trunk girth, percentage of the vertical canopy cover, and others). The method presented below is a photographic technique which gives good estimation of the effect of the vegetation cover on the radiation balance of a surface beneath a canopy.

Measurement may be done with a specialized wide angle camera or with a conventional camera with a wide angle lens. The camera used in the present study is a Russian-made Horizont, panoramic (Zenith) camera, which is capable of photographing an angle of 120 degrees with visible, but acceptably small, distortion.

Figure la illustrates the way in which the camera functions. The lens is mounted in a rotation turret which scans 120 degrees. The light beam passing through the lens rotates across the film, which is held tightly against the back plate of the turret. The back plate of the turret has the same radius of curvature as the turret. This scanning provides a near linear image of the portion of the hemisphere seen by the lens. Distortion is at a minimum, as can be seen from Figure lb. The resulting negative is 6 cm long on 35 mm film.

Since the camera sees only 120 degrees as a maximum, two photos must be taken in order to obtain a complete cross-section of the hemisphere. To achieve this the camera is mounted on a block of wood, which makes an angle of 30 degrees with the horizontal. This inclination allows the camera to see from the ground surface to 120 degrees. The block is then rotated





↑ Fig. 3: Rate of canopy-thinning with altitude under different vegetation.

← Fig. 2: Proportion of sky seen at different angular intervals.

180 degrees, and a second photo is taken from 180 degrees to 60 degrees (Fig. 1c). Several pairs of photographs can be taken in a vegetation unit to obtain an adequate sample from the circumference of the hemisphere. An example of a pair of photographs is shown in Figure 1d.

The procedure described above may be carried out with a camera having a smaller angle of view, but it requires that the block have several angles cut into it so that a cross-section of the hemisphere can be obtained. Alternatively, this could be done by mounting the camera on a short tripod which would allow the camera to be rotated through several angular increments. The prints could then be pieced together.

Analysis of the Photos

Prints of a standard size (in the present case, 12 cm by 7.3 cm) are produced. The print is divided into approximately equal vertical angular intervals representing a small portion of the hemisphere, e.g. 2 degrees, laying a fine grid over the photographs. A point count is taken of the number of points falling on open sky along any one horizontal line of the grid and this number of points is expressed as a decimal fraction of the total number of points on the line. This is the decimal fraction of the sky seen in that angular interval.

In order to characterize the shading properties of the vegetation, the decimal fraction of the sky seen in an interval was plotted against the mean angular distance of the interval from the ground surface. The technique was used in both boreal forest and regenerating burnt areas. For such sites, the proportion of sky on any small portion of the hemisphere is a linear function of angular altitude (Fig. 2) and can be expressed as:

$$s = a + b\gamma \qquad (1)$$

where s = proportion of sky "seen" by the ground in some angular interval;

γ = the mean angular elevation of the interval and ranges from 0 to the angular height of the vegetation;

a,b=emperical constants, characteristic of the vegetation type, especially of its density, height, and shape.

The parameter <u>s</u> indicates the extent to which direct solar radiation can penetrate the canopy when the sun is at a given elevation, γ, less than the angular height of the vegetation. When γ exceeds the angular height of the vegetation, s = 1.00. Alternatively, <u>s</u> may be thought of as the proportion of the time during which the direct solar beam is not intercepted by trees, while it is moving through some small increment of altitude around γ.

The coefficients, a and b, vary with the characteristics of the vegetation. The value of a represents the sun altitude at which the ground is completely shaded from the direct solar beam. It can be thought of as the skyline, or effective horizon for the particular vegetation type. If a is small, for example, the vegetation is open almost down to the ground surface. The value of b represents the rate at which the canopy thins with altitude. In a tall, dense canopy, b is small, whereas in a regenerating burnt area, it is very high (Fig. 3). The sampling characteristics of these a and b values within various boreal forest communities are currently being studied.

It is possible that in other vegetation communities a different relationship between altitude and shading may prevail. For other regions, semilogarithmic or polynomial functions may provide a better fit to field data. In the extreme case, (Reifsnyder and Lull, 1965), in which the vegetation can be described as a solid cylinder, equation (1) would be a step function in which:

$$\begin{aligned} s &= 0 \quad \text{for } \gamma < \text{arc tan } 2\ H/D \\ s &= 1.0 \text{ for } \gamma > \text{arc tan } 2\ H/D \end{aligned} \qquad (2)$$

where H = average tree height; D = average distance between trees.

The preceding discussion involves the shading effects of the canopy with respect to direct solar radiation. In the energy balance, one must also consider the entry of diffuse solar radiation and the exchanges of longwave radiation between snow surface, trees, and sky. All of these components are non-directional. They occur over the relevant portion of the whole hemisphere.

Diffuse radiation reaches the ground from that portion of the sky not obscured by trees. This sky view factor is obtained by weighting the proportion of sky seen at each angular elevation by the fraction of the total area of hemisphere at that elevation. Thus, the view factor (P) can be defined as:

$$P = \sum_{\gamma=0}^{80^\circ} s_{\gamma+5^\circ} \int_{\gamma}^{\gamma+10^\circ} \cos\gamma \cdot d\gamma \qquad (3)$$

which reduces to:

$$P = \sum_{\gamma=0}^{80^\circ} s_{\gamma=5^\circ} \{\sin(\gamma+10^\circ) - \sin(\gamma)\} \tag{4}$$

for $\gamma = 0^\circ, 10, 20\ldots, 80^\circ$.

When γ is greater than the angular height of the trees, $s_\gamma = 1.0$.

The exchange of longwave radiation between sky and ground surface also takes place over the proportion of the sky represented by P. The exchange of longwave radiation between trees and ground surface occurs over the area (1-P).

A complete radiation balance for a forest floor, therefore, can be expressed as:

$$\{Q(\gamma)s(\gamma) + q(\gamma)P\}\{1 - \alpha(\gamma)\} + L{\downarrow}P + T{\downarrow}(1-P) - S{\uparrow} = N \tag{5}$$

where $Q(\gamma)$ = direct beam solar radiation, a function of sun altitude;
$S(\gamma)$ = proportion of the sky "seen" by the ground over a small angular interval at the altitude of the sun;
$q(\gamma)$ = diffuse shortwave radiation, a function of sun altitude;
P = view factor;
$\alpha(\gamma)$ = albedo of the snow, a function of sun altitude;
$L{\downarrow}$ = longwave radiation from sky to snow;
$T{\downarrow}$ = longwave radiation from trees to snow;
$S{\uparrow}$ = longwave radiation from snow to atmosphere and trees;
N = net radiation at the forest floor, snow pack surface, or other elevation beneath the canopy.

It is possible to write a similar equation for any other level under the canopy, such as the top of a shrub layer.

The technique described has been used to calculate the energy balance of the snow pack in a boreal forest. There may be other applications, such as calculating the growing season input of radiation to the under storey vegetation on a forest floor. Other uses can be envisaged in forestry. In areas where forest vegetation is being removed, the increased insolation received by streams causes an important increase in the water temperature. To predict this increased temperature, energy balance calculations are used (Brown, 1970). One way of reducing or eliminating the change is to leave buffer zones of vegetation along the stream channel. There is some problem in deciding how wide the buffer zone should be, and which kind of vegetation will shade a stream of a given width and general orientation in a certain latitude. The photographic technique that has been discussed here permits a modification of the energy balance to incorporate the effects of these variables, thus allowing design calculations to be made from some simple photographic measurements.

Acknowledgements

This work was carried out with the assistance of grants from the National Research Council and the Department of the Environment.

REFERENCES

Anderson, M.C., 1964: "Light Relations of Terrestrial Plant Communities and their Measurement", Biological Reviews, Vol. 39, pp. 425-486.

Evans, G.C., and D.E. Coombe, 1959: "Hemispherical and Woodland Canopy Photography and the Light Climate", Journal of Ecology, Vol. 47, pp. 103-113.

Johnson, P.L. and T.C. Vogel, 1968: Evaluation of Forest Canopies by Photography, U.S. Army, Cold Regions Research and Engineering Lab Research Dept. 253.

Monsi, M. and T. Saeki, 1953: "Uber den Lichtfactor in den Pflanzenge svell schatten und seine Bedeutung für die Stoffproduction", Japanese Journal of Botany, Vol. 14, pp. 22-52, in Anderson (1964).

Reifsnyder, W.E. and W.L. Lull, 1965: "Radiant Energy in Relation to Forests", Technical Bulletin No. 1344, U.S. Dept. of Agriculture, Forest Service.

ENERGY BALANCE COMPUTATIONS OF SNOWMELT RUNOFF IN THE SUBARCTIC

by

A.G. Price, T.Dunne and J.A.V. Burns*

The energy balance method is well-known in hydrometeorology. It is the method whereby all the heat exchanges across a surface are assessed or measured, enabling the computation of such factors as melt, evaporation, temperature or heat storage. When applied to snow the energy balance method requires the definition of the five major heat fluxes between the snowpack and its environment. The total energy balance equation may thus be written:

$$Hm = Hg + Hp + He + Hc + Hr + H\theta \qquad (1)$$

where Hm = heat available for melting of snow or freezing of liquid water, Hg = ground heat flux, Hp = heat supplied by rainfall, He = heat from the turbulent transfer of water vapor (latent heat transfer), Hc = heat from turbulent transfer of sensible heat (sensible heat transfer), Hr = heat from radiation, and $H\theta$ = change in heat storage. Although this equation does not define the heat balance completely, the terms expressed embrace the great majority of heat exchanges over the snowpack. A consideration of all the individual terms will allow further simplification.

The ground heat flux (Hg) except in unusual circumstances is a negligible proportion of the total heat flow. It is generally assigned a small positive constant amount during the whole melt (U.S. Army, 1956). This amount is usually about 2mm/dy. In the Schefferville area, field observations and measurement show that this heat flow is zero during the main melt. It has, therefore, not been included in the calculations of

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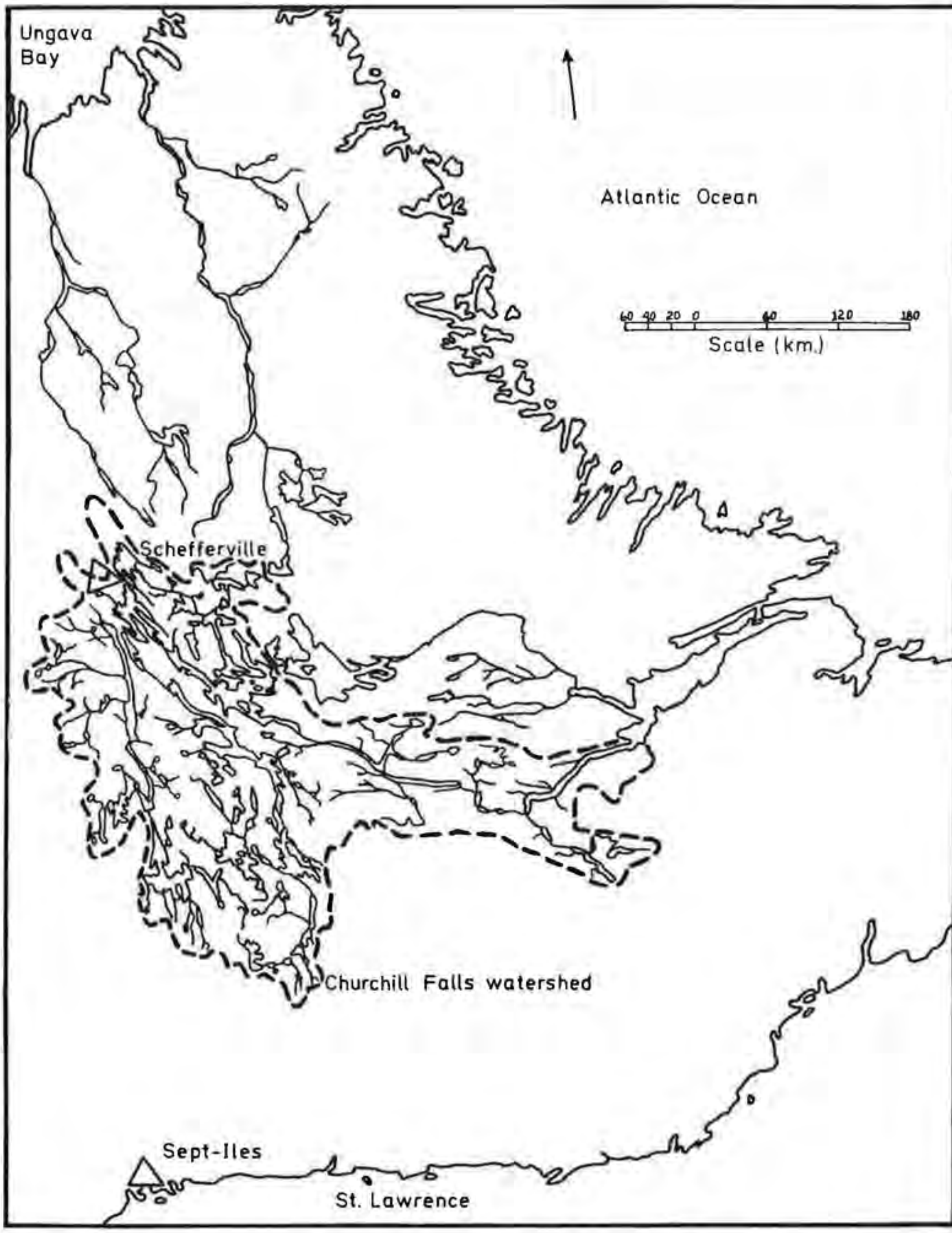

Fig. 1: Location Map - The Schefferville Area.

melt.

Heat storage change ($H\theta$) for a melting isothermal pack is zero ($H\theta=0$).

Heat supplied by rainfall (Hp) must be considered when rainfall at a temperature greater than 0°C supplies some heat for the melting of snow. This amount of heat can be calculated from the amount of precipitation and the wet-bulb temperature, the wet-bulb temperature being equated to rain temperature. The flux (Hp) occurs only during rainfall and is very small relative to the other components. Throughout the melt period discussed here this term did not enter the energy balance equation.

Thus the original energy balance equation can be simplified to three terms: Hc, He, and Hr.

The turbulent heat fluxes (Hc and He) are caused by turbulent transfer of sensible heat and moisture in the lower atmosphere over the snowpack. Estimation of these two fluxes is usually achieved by using equations of the form:

$$He = f(Vb, \{e_a - e_s\}) \tag{2a}$$

$$Hc = f(Vb, \{T_a - T_s\}) \tag{2b}$$

where Vb = windspeed, e_a = vapor pressure of the air, e_s = vapor pressure of the snow surface, T_a = air temperature, and T_s = surface temperature of the snow. The most commonly used equations are those of the U.S. Army Corps of Engineers (USACE) and Sverdrup, representing the combination of energy balance and aerodynamic methods (USACE), and the pure aerodynamic method (Sverdrup). As an example the USACE equations are:

$$He = k'(za \cdot zb)^{-1/n}(e_a - e_s)Vb \tag{3a}$$

$$Hc = k''(za \cdot zb)^{-1/n}(T_a - T_s)Vb \tag{3b}$$

where za and zb are the heights at which the measurement of air temperature (T_a), vapour pressure (e_a) and windspeed (Vb) have been taken. These particular equations assume that the eddy diffusivity for momentum is the same as that for water vapor, and that for heat. In addition they assume that

$$Qz = Q_1 . z^{1/n} \tag{4}$$

where Qz = value of a property of the atmosphere at height z, and Q_1 is the value at any reference level. The quantity 1/n is .16 for adiabatic (unstratified) conditions (Brutsaert and Yeh, 1971; Sellers, 1965). The adoption of a value of 1/n of .16 is the more severely restrictive of the two assumptions. Neutral conditions over the snowpack are certainly the exception rather than the rule, and therefore these equations will be seriously in error under certain conditions. This will be discussed later in the paper. The empirical coefficients k' and k" were derived from lysimeter snowmelt work carried out by USACE.

Spatial Variation of the Energy Balance

The spatial variation of the three major heat sources, He, Hc and Hr, has not been the subject of extensive study, the work of Hendrick (1970) being an important exception. Variation of heat input is caused by differences in vegetative cover and by topography.

The effect of topography upon the radiative heat input depends on the aspect and inclination of the ground surface. The resulting variation in the radiation conditions may be geometrically computed using the method developed by Garnier and Ohmura (1969). Their mapping program computes net radiation for slopes of specified inclination and aspect from measured or estimated values on a horizontal surface.

Variations in topography will also cause spatial variation in the terms of turbulent exchange. Amounts of heat transmitted to or from the snow are a linear function of windspeed. Thus, variations in windspeed at any one level caused by differences of elevation and exposure will cause changes in the magnitude of the He and Hc terms. This variation of windspeed, however, was not modelled because of lack of time and because of insufficient instrumentation. In any case under normal conditions the turbulent heat exchange is much less than Hr.

The most conspicuous variation in all the terms is caused by differing vegetative covers, particularly the difference between forested and non-forested areas. Forest considerably modifies the radiation balance at the snow surface. A model has been developed which describes the modification of the radiation balance by tree cover of various types. The model predicts lower peak values of Hr in the forest as compared to that in the open, and a smaller night-time loss of longwave radiation from the snowpack in the forest as compared with the open. Both of these general tendencies are seen in measured values taken in the forest and in the open. Incoming shortwave radiation is depleted by the canopy, whereas incoming longwave radiation is increased by the longwave income from the forest canopy. Differing types of forest cover have different effects upon the radiation balance, and the model incorporates such variations.

The presence of the forest also modifies the turbulent exchange terms Hc and He. It is thought that although the windspeed at any level is less in the forest than in the open, the same general profile form exists in the forest as in the open, i.e. the normal semilogarithmic profile:

$$u = u^*/k \quad \ln(z/z_o) \tag{5}$$

where $u^* = (\tau/p)^{\frac{1}{2}}$ (Sellers, 1965, p. 148).

Although appealing to interference by the trees in order to account for reduced general wind velocities, it is necessary to assume that this interference does not change the wind profile form in any consistent fashion in the space between the trees.

As noted earlier, thermal stratification of the air layers above the snow has substantial effects on the turbulent exchange of heat and moisture to and from the snow surface. If the wind speed increment between any two levels is known, and also the temperature increment between the same two levels, then a stability factor Ri can be written:

$$Ri = \frac{g\Delta t\Delta z}{(\Delta u)^2 T} \tag{6}$$

where g =acceleration due to gravity cm sec^{-2}, Δt = temperature increment °C, Δz = height increment/cm, Δu = windspeed increment cm sec^{-1}, and T = mean temperature in the interval (°K). Using this term, corrections can be made for the damping effects of stratified air on the turbulent exchanges. In neutral conditions $\Delta t = 0$, and $Ri = 0$. Alternatively, a diabatic profile may be written (Swinbank, 1964):

$$\frac{u}{\ln(\frac{z}{z_o})} = \frac{u^*}{K} \quad \ln(\frac{z}{z_o})^{(1-\beta)} \tag{7}$$

where $(1-\beta)$ is a stability parameter, and when $\beta = 1.0$, the profile form reverts to the neutral case (Sellers, 1965).

Using these formulations, it is possible to compute the turbulent fluxes under varying conditions of stability. However, two points should be made. First, Hc + He is normally a very small term, and so errors are relatively unimportant. Secondly, under conditions where Hc and He are large, Ri is likely to be approximately equal to zero. That is, when the fluxes are large, the assumption of neutrality is less dubious than under conditions where the fluxes are small. In the latter case, amounts of heat are so small that errors are insignificant. At the peak melt period under high radiation conditions, the term Hc + He was less than 5% of the total melt. Considerations of the stability conditions at the time

TABLE ONE

Characteristics of the Experimental Sites

Site	Area (m^2)	Approximate Length	Mean Gradiant	Aspect (°)
Tundra				
A	30,250 sq. ft.	280'	4°25'	242°
B	14,370 sq. ft.	160'	9°00'	271°
C	19,130 sq. ft.	120'	5°05'	058°
Forest				
D	25,930 sq. ft.	280'	7°10'	062°
E	19,400 sq. ft.	180'	15°05'	215°
F	19,620 sq. ft.	200'	5°30'	013°
G	18,080 sq. ft.	250'	7°05'	244°

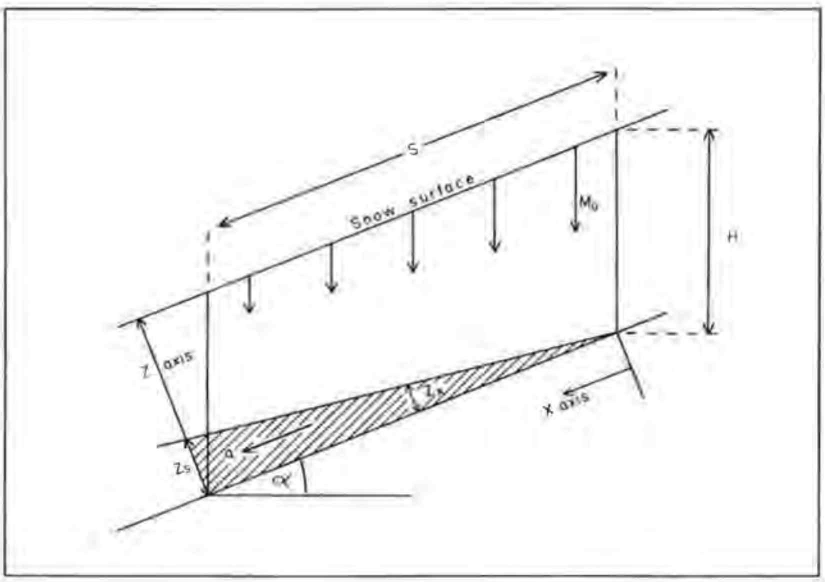

Fig. 2; Diagram to illustrate the Mass Flux of Water through a Snow pack.
(after de Quervain)

suggest that turbulent exchange was so damped in this period (by stable stratification) that fluxes must have been approximately zero. Application of the standard equations gave an extremely low positive figure. It may also be noted at this point that the power law form of the wind velocity profile under neutral conditions yields a value of .166 for $1/n$ (eq. 4). Over the whole period, field measurements of the wind profile yielded a value $1/n = 0.7$. This suggests that overall, conditions of very high stability occur over the snowpack in the study area.

The Experimental Area

The study is being conducted near Schefferville, Quebec. Schefferville stands at the divide between the Ungava Bay drainage system and the Hamilton-Ashuanipi (Churchill Falls) system (Fig.1). The topography in the Schefferville area is generally of a ridge-and-valley type, being strongly controlled by structure. The climate is typical of the Eastern Canadian Subarctic and the thaw generally commences in May and continues through June. The annual snowfall in the area, expressed in terms of water equivalent, is about 35cm with a record high of 41cm in 1969-70, and a record low of 27cm of annual precipitation in the area. The region is covered by two dominant vegetation types: tundra and "open lichen woodland," a spruce forest with low crown closure (approximately 25%), little understory and a thick lichen mat.

The Experimental Sites

Three experimental plots are located in the tundra and four in the spruce forest. The aspect, gradient and dimensions of each of the sites are shown in Table One.

The sites are located in two clusters and at each cluster the following meteorological variables were monitored:*

1. Incoming shortwave radiation. A single bimetallic actinograph was placed on the tundra, and in the forest five Lintronic solarimeters were operated in series.
2. Albedo. A pair of Lintronic solarimeters facing up and down was located in the forest. In the open, occasional measurements of albedo were made.
3. Net allwave radiation. Five net allwave radiometers (CSIRO, Funk-type instruments) were operated in series in the forest. On the tundra net allwave radiation is estimated from shortwave radiation.

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* All data on radiation used in this study were supplied by R.G. Wilson and D. Petzold whose assistance is gratefully acknowledged.

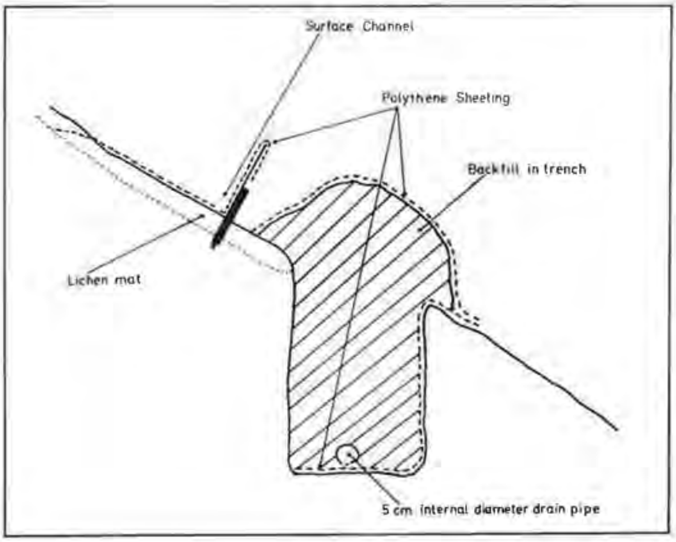

Fig. 3: A Diagram showing the Construction of Trenches to measure Subsurface Flow.

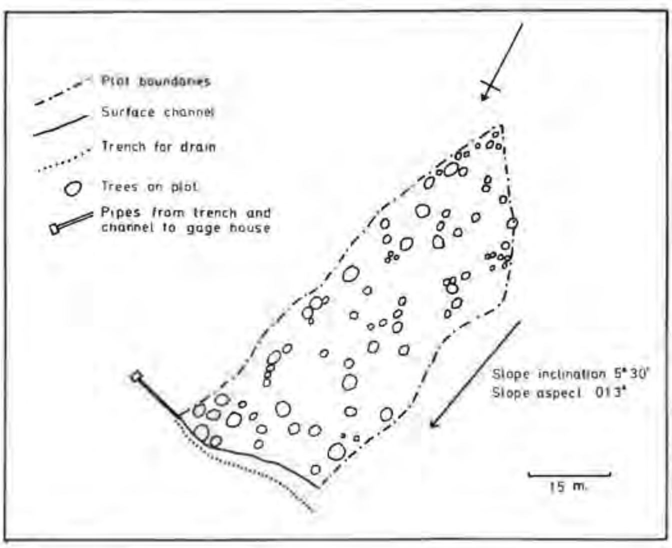

Fig. 4: A Map of Plot "F" (Forest Site).

4. Windspeed. Hourly mean windspeeds at two meters were measured at both clusters with standard cup anemometers. At the forest sites the form of the wind profile was measured at a mast holding four anemometers at heights of 15cm, 50cm, 1m, and 20m respectively.

5. Atmospheric vapour pressure and temperature. Hourly readings of temperature and vapour pressure were made at both clusters using an Assman aspirated psychrometer, and a hygrothermograph to provide a continuous record.

At each of the seven sites the following hydrologic measurements were made:

1. Amount of snow. On each plot approximately 20 stakes were used for daily measurements of snow depth. Surveys of snow density were made with an Adirondack snow-sampler on alternate days and used to convert the snow depths to water equivalent.

2. Runoff. The runoff from each plot was intercepted and recorded continuously at the lower end of each plot. Surface runoff was intercepted in a channel, and subsurface flow by a tile at a depth of 3 to 4 feet (Fig. 3). Severe leakage occurred from some of the surface runoff collectors, and they have been replaced by a more reliable system for use during the second field season.

Figure 4 shows the characteristics of a typical plot in the forest.

Results

The data are still being analysed and another field season is planned, but the following is a sampling of the kind of results obtained to date.

Using the techniques described at the beginning of this paper, an hourly energy balance can be computed for each site. Figure 5 presents such results. The radiation values shown have not been corrected for slope, but measured values of radiation on a horizontal surface have been used for both the open and forested sites. In each diagram, the hydrograph of surface runoff is included for comparison of its timing with that of the computed melt. Subsurface runoff did not occur until after the snow had melted completely. The ground was frozen to great depths and was completely impermeable.

Analysis shows that the variations in timing of the imputs is insufficient to explain all the differences in the hydrograph lags behind input. The other variables which explain the different responses shown in the diagrams are those describing the shape of the slope and the properties of the snow: that is, the hydraulic variables. If we can define the surface fluxes using the energy balance, then models have already been developed which will describe the passage of a wave of mass flux through the snowpack (Colbeck, 1970). An adaptation of this model to two-dimensional flow

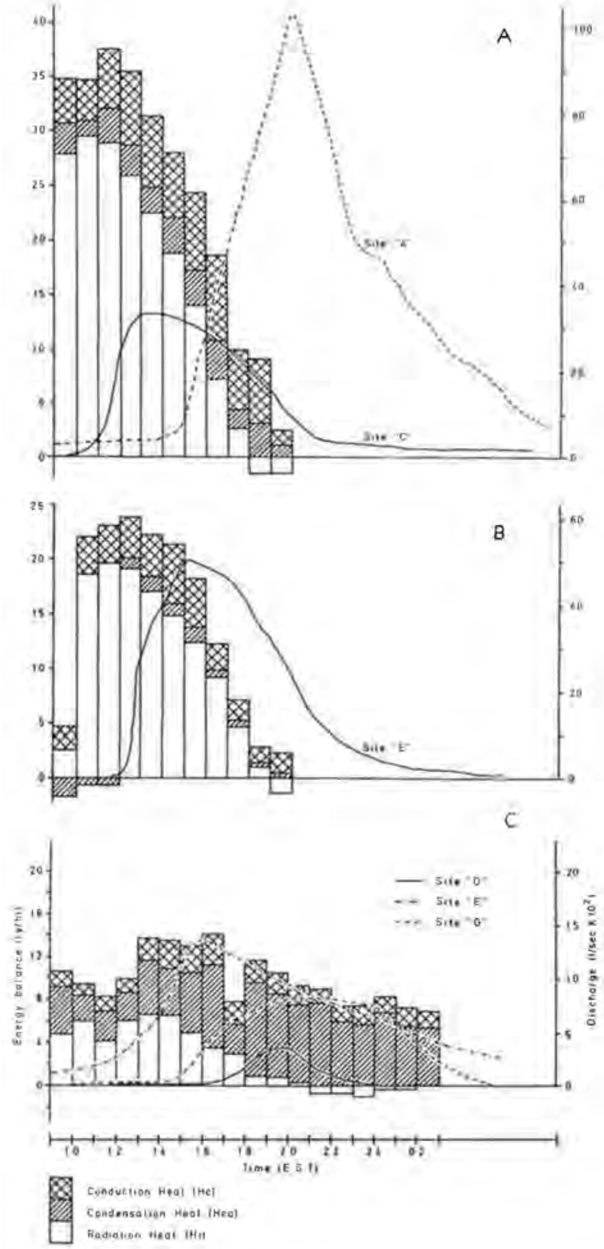

Fig. 5: The Energy Balance & Runoff at different sites.
A: Tundra, May 16, 1972.
B: Site "E", aspect 215°, Forest, May 16, 1972.
C: Sites "D","G", and "E", Forest, June 3, 1972.

is currently being developed. The steady-state situation is very simple to define (de Quervain, 1972) and is illustrated in Figure 2.

Using Darcy's law we can thus write:

$$Q(x) = b \cdot z(x) \cdot Kw \sin\alpha \quad (8)$$

where $Q(x)$ = mass flux of water at distance S from the divide, b = plot width, $z(x)$ = thickness of saturated layer at distance S from the divide, Kw = saturated conductivity of snow and α = angle of slope.

If we can define the necessary variables, then we can predict Q(x), which is what is being measured in the field. The problem essentially is solving the equations of unsaturated flow, and making the model time-dependent.

The above is a physical hydraulic model. It should also be possible to model the flow through the snow and down the slope in an empirical fashion, by using techniques such as the instantaneous unit hydrograph (Viessman, 1968). The output might be generated from the input by the application of a transfer function. A further possibility is the application of stochastic modelling techniques to both the input and the output data, although superficially this seems less promising.

Lastly, time-series analysis obviously has great potential in this particular analysis, since essentially an attempt is being made to correlate two time-series.

Acknowledgement

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REFERENCES

Brutsaert, W. and Gour-Tsyh Yeh, 1970: "A Power-law Function for Turbulent Transfer Computation", Water Res. Research, Vol.6, No.2, pp. 1387-1391.

Colbeck, S., 1971: One-dimensional Water Flow through Snow, U.S. Corps of Engineers, Cold Region Research and Engineering Lab.

De Quervain, M.R., 1972: "Snow Structure, Heat and Mass Flux through Snow", International Symposia on the Role of Snow and Ice in Hydrology, UNESCO.

Garnier, B.J., and A. Ohmura, 1968: "Computation and Mapping of the Short-wave Radiation on a Slope", Jnl. Appl. Met., Vol.2, pp. 796-800.

Hendrick, R.L., B.D. Filgate and W.W. Adams, 1971: "Application of Environmental Analysis to Watershed Snowmelt", Jnl. Appl. Met., pp. 418-429.

Sellers, 1965: Physical Climatology, Chicago University Press.

Sverdrup, H.V., 1936: "The Eddy-conductivity of the Air over a Smooth Snow Field", Geofys. Publ., Vol.11, No. 7.

Swinbank, W.C., 1964: "The Exponential Wind Profile", Quart.Journ.Roy. Met.Soc., Vol. 90, pp. 119-135.

U.S. Army Corps of Engineers, 1956: Snow Hydrology, N. Pacific division, USACE, Portland, Oregon.

SNOWMELT, RUNOFF, AND BREAKUP IN THE COLVILLE RIVER DELTA, 1971

by

Jeffrey S. Peake and H.J. Walker*

Rivers which flow into the Arctic Ocean yet which originate outside of the Arctic, such as the Mackenzie of Canada and the Yenisey of Siberia, continue to flow during winter even though their surface may be frozen to depths of two meters. On the other hand, rivers which have their entire drainage basin within the zone of continuous permafrost virtually cease flowing during winter because nearly all of their surface and ground water is locked in a frozen state for up to eight months of each year. The Colville, longest river on the North Slope of Alaska, is an excellent example of such a river.

The Colville originates in the Brooks Range and crosses the Arctic Foothills and Coastal Plain Provinces before entering the Beaufort Sea (Fig.1). The delta it has produced is 610 km^2 in area and is actively growing today. It consists of several distributaries, numerous lakes, large sandbars and mudflats which are flooded each spring, partially stabilized dunes which form on the west bank of many of the river's channels, and the almost ubiquitous ice-wedge polygon.

During 1971 a study was made of snowmelt, runoff, and breakup in the Colville system. The research, which concentrated on how these variables affected the flooding within both the subaerial and subaqueous portions of the delta, began in late April and lasted beyond the snowmelt period. Observations were made and data were gathered at the Naval Arctic Research Laboratory camp at the head of the delta. With the aid of snowmobiles, boats, planes and a helicopter, it was possible to record processes in other areas of the delta for varying times.

Prior to the initiation of snowmelt, which along the Arctic coast usually begins in mid-May, the Colville Delta is characterized by:

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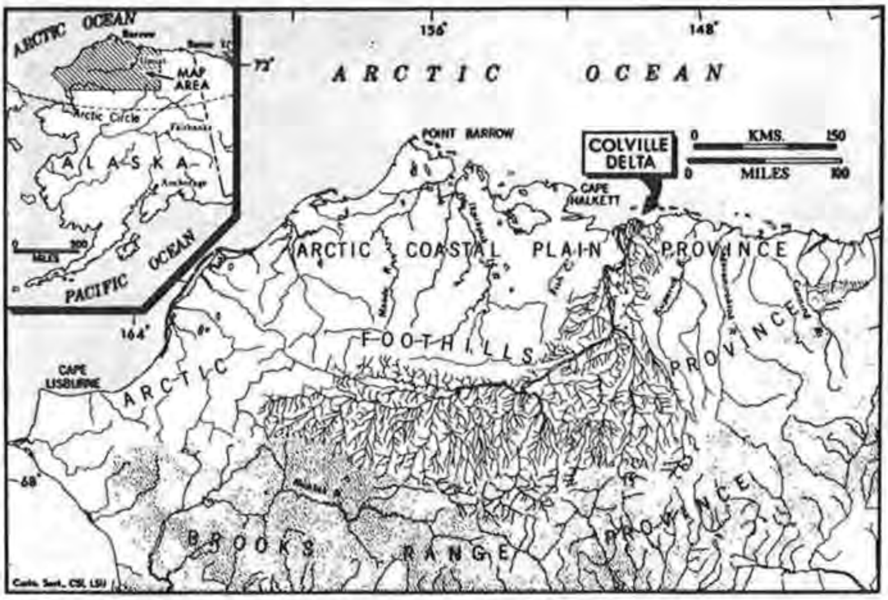

Fig. 1: Arctic Alaska and the Colville River system.

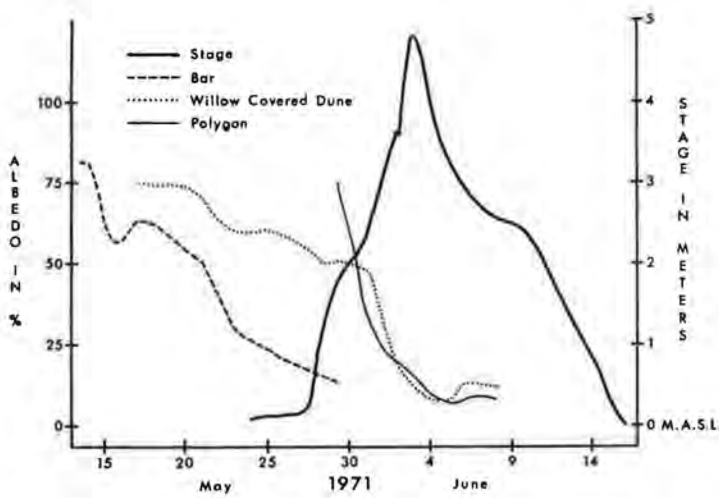

Fig. 2: Decline in Albedo over Selected Surfaces and River Stage at the Head of the Colville Delta, 1971.

1) a two meter thick ice cover on all deep bodies of water including lakes, distributaries, and the ocean; 2) bottom-fast ice in water bodies which were less than about two meters deep at the time of freezeup; 3) the river (including its ice cover) at a low stage; 4) the presence of sea-water within the sub-ice portion of the delta's distributaries; and 5) a minimal amount of hydrologic, oceanographic, and morphologic activity.

All surfaces are overlain by a snow cover which averages about 35 cm in thickness. Generally the snow cover has a relatively uniform upper surface, although its depth varies, depending mainly upon the unevenness of the tundra surface. Although wind occasionally exposes lake ice, sandbars, and dune crests, it is also responsible for the formation of drifts several meters thick against steep riverbanks and in dune depressions.

The snow has an average density of between 0.3 and 0.4 gm/cm^3 and an albedo of between 80 and 90 percent during the premelt period. Such high reflectivity delays melting of most of the snowpack until late May and early June. The nature and timing of runoff and breakup are affected not only by the amount of snow but also by the rate at which it melts. The time of melting varies with the depth of snow and may be affected by the nature of the microrelief (such as grass tussocks). Several areas, mostly those where there was little snow to begin with, such as dune crests and river bars, are rapidly depleted of their snow cover. As the albedo of these areas diminishes, there is an increase in the net radiation absorbed at the surface. Much of this energy goes into thawing the active layer and into creating rather high surface temperatures (especially on the dune crests, from which meltwater drains rapidly). Such high surface temperatures, although localized, warm the air near the ground and appear to aid considerably in the melting of adjacent snow patches. In 1971, the time when most melting occurred is suggested by the change in albedo over different surfaces (Fig.2).

In late May, meltwater begins to accumulate on the river ice. The initial water is from the melting of the snow present on the river ice, on sandbars and mudflats adjacent to the river, and from snowdrifts along riverbanks. After a few days the meltwater on the river ice begins to flow downstream. Although the initial accumulation of melt-water is slow, once the snow on the tundra begins to melt and the melt-water from upstream begins to reach the delta, the stage rises rapidly (Fig. 2). Pre-breakup floodwater flows over bottom-fast ice and beneath

Fig. 3: Looking seaward over the front of the Colville delta during prebreak-up flooding. Note floodwater on tundra and deterioration of snow cover (lower right). River ice is floating in place.

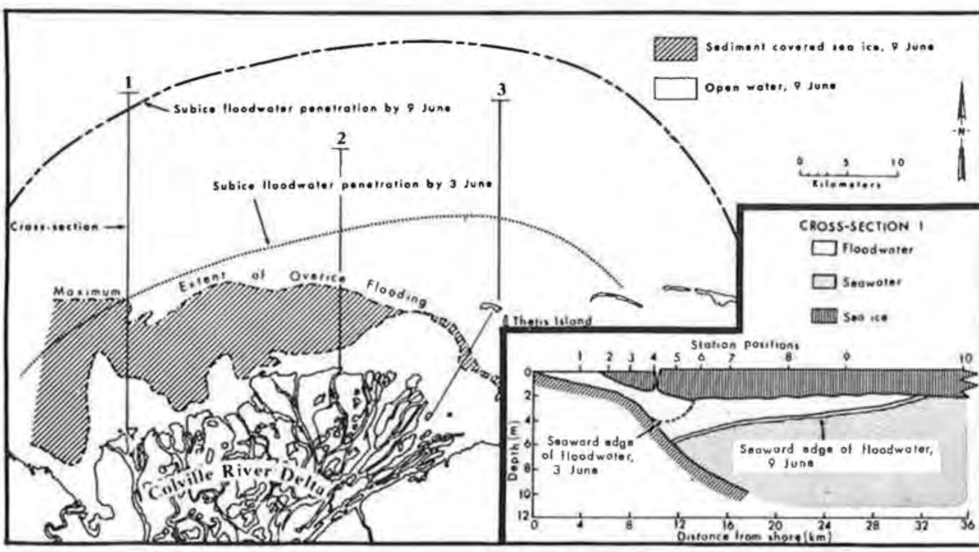

Fig. 4: Extent of flooding above and beneath the sea ice in 1971. Cross-section shows seaward progression of freshwater wedge between June 3 and June 9.

the ice which was not bottom-fast (Fig. 3), melting the ice in the process.

Breakup, here considered to be the period during which the ice is removed from the channel, is of short duration, usually lasting less than two days at any one point along the river. The time of breakup at various points in the delta differs. In 1971 it occurred nearly three days later at the delta front than at the head of the delta, 40-45 km upstream. By this time (June 3) nearly all of the snow cover in the delta had been removed, whereas the tundra surface both east and west of the delta was still nearly 50 percent snow covered.

At the height of the flood in 1971, nearly 65 percent of the delta was covered with floodwater. Such extensive coverage aids in the removal of snow directly or indirectly in several ways: 1) it undercuts snowdrifts along the riverbanks and carries snowblocks seaward; 2) it melts the snow cover over which it flows; 3) it rapidly lowers the albedo of the surfaces over which it flows; and 4) it increases the thermal energy within the delta as a whole. During the period of breakup water temperatures of up to 5°C were recorded.

As floodwater leaves the river channels, some of it flows under the ice at the front of the delta. Most of it, however, flows on top of the sea ice (Fig. 3) to a distance of several km offshore where it drains through cracks and holes associated with pressure ridges. As the floodwater flows out over the sea ice, it deposits much of the sediment it has carried from the river. In 1971 deposits up to 17 cm thick were measured on the ice north of the main channel. The floodwater reduces the albedo of the snow-covered sea ice rapidly; although the floodwater soon drains from the ice surface, the reduced albedo remains low because of the sediment cover which has been left.

In 1971 floodwater reached its maximum distance on top of the sea ice in about three days. The flow beneath the ice progressed as a freshwater wedge at an average rate of about 5 cm/sec, reaching distances of 30 to 40 km within 10 days after floodwater began to flow into the ocean (Fig. 4). The freshwater-seawater interface was so distinct that volumetric calculations of seawater displacement beneath the sea ice were possible.

The duration and nature of post-breakup flooding reflects the amount of snow left after breakup and the rate at which it melts. During most years this period ends in mid-June and is longer than the other two periods, i.e. pre-breakup and breakup flooding. By the time the river drops to a typical summer level, virtually all of the snow is gone. However, there is still surface ice left in the delta. Large lakes which were not flooded by the river may retain their ice cover until late June. There are also some protected areas where large snowdrifts may persist for several weeks after most of the surrounding snow has disappeared. Thus, even that late in the high sun period, a small portion of the delta surface retains relatively high albedos.

It is quite obvious that the amount and rate at which the snow melts has great influence on the nature and timing of breakup and flooding within the delta. Although less obvious, it is nonetheless just as true that the reverse is equally valid, namely that the nature and extent of breakup and flooding is important in the disappearance of the snow within the delta.

Acknowledgement

The research reported on here was supported by the Arctic Program and Geography Program, Office of Naval Research, under Contract N00014-69-A-0211-0003, Project NR 388 002, with Coastal Studies Institute, Louisiana State University. Logistic support was provided by the Naval Arctic Research Laboratory, Barrow, Alaska.

A NOTE ON DEGREE DAY ACCUMULATION AND GLACIER RUNOFF SIMULATION

by

Sam I. Outcalt*

In a short note Andrews et al (1971) have exposed the dangers involved in the use of "cumulative data" in glaciological studies. Their note demonstrates that extremely high correlations may be obtained between uncorrelated data vectors when the data elements result from an accumulation process. This fact was not given proper attention in the monograph by Outcalt and MacPhail (1965) due to a lack of sophistication on the part of the senior author.

However, with this note Andrews et al (1971), pose a question which transcends their discussion of the statistical manipulation of degree-day data: namely, the physical reality of the degree-day concept and the validity of the application of this statistic in geophysical analysis.

The Physical Interpretation of Degree-Days

Degree-days have the dimensions of a temperature*time product and the statistic is the integral beneath a thermograph trace. If the freezing point is selected as the reference point in glaciological studies both positive and negative degree-day statistics are possible. The statistic (time*temperature integral) is physically derived from the integration of the differential equation equating the surfaceward heat flow and fusion at a descending freezing or thaw plane in a lake or soil. In this context the statistic springs directly from the energy conservation rule and can hardly be considered a handy choice (Yong and Warkentin, 1966).

If the air temperature at screen elevation is used as an index of the total energy available for ablation on a nearby glacier or snow surface, theory and observation are often at odds. This author has

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frequently encountered conditions, in environments ranging from the High Arctic to the Colorado Rockies, when surface melting was observed at screen level temperatures well below the ice point. These conditions are most commonly encountered when the surrounding terrain is largely ice-snow covered. The conclusion as to the validity of this application of the degree-day statistic is contained in the fact that the air is heated primarily by the surface and advective effects diminish away from the land margin. In most studies of temperate mountain glaciers the weather screens are located at a nearby valley village or at the campsite on land near the glacier margin. The diurnal thermal regime observed at such a weather screen would follow the surface regime with some amplitude constriction and phase lag. There is no reason to believe that the time*temperature integral at such a site is not strongly correlated with the energy absorbed over <u>that surface</u>. However, on the nearby glacier the entire energy transfer environment is vastly different due to a higher albedo and the diminished magnitude of net thermal radiant and turbulent transfer by fusion limiting the maximum surface temperature to the ice point.

However, statistical studies normally indicate that total glacier runoff and degree-days accumulated over land are highly correlated. In fact, both of these are "carts" being drawn primarily by the "horse" of incoming solar radiation as modified by airmass conditions (wind, temperature, humidity and turbidity) and the local surface geography (albedo, roughness, wetness and thermal properties). Thus, in the glaciological case, correlation is developed between a point (weather screen) and a snow-watershed basin. Both the point and the basin by definition have unique and radically different geographic boundary conditions which govern the nature of the complex transfer function which related incoming solar radiation to the surface thermal regime and energy transfer environment. This situation is severely in contrast with the thermal soil mechanics application of the degree-day statistic where accumulated air or surface temperature is used as an index of substrate enthlopy <u>at that point</u>. Lastly, the philosophical justification for the statistical treatment of serial thermal data is weak due to the failure of the data to meet the criteria of randomness, independent selection and homogeneity which are basic to parametric regression analysis. This condition is produced by strong serial correlation in time-dependent data and the problem of defining a homogeneous population spatially and temporally (Bruce and Clark, 1966).

The Information Content of Regression Parameters

Due to the weak and non-isomorphic match between conditions in hydrometeorology and the foundations of a mathematical scheme called parametric statistics, "practical men" fall back on "reliable prediction" as a means of evaluation. It has been the practical experience of operating hydrologists that these methods are both necessary and sufficient for their forecasting needs due to financial and analytical restrictions. However, these statisticians wishing to increase glacier runoff would have only the option (in a theoretical framework cast by regression parameters) of building a fire under the local weather screen whereas a student of the surface energy transfer environment might choose to lampblack the glacier surface. In an attempt to formalize these notions the author constructed a synthetic glacier-runoff simulator within the bowels of the Michigan Computer System.

The Temperate Glacier Runoff Simulator

This simulator was developed as a modification of a general surface climate simulator which is discussed elsewhere (Outcalt, 1971). The glacier-runoff simulator is designed to approximate the energy transfer of a temperate glacier after it has become seasonally isothermal and the winter cold wave has been dissipated. The general simulator was modified to make the glacier surface a soil heat sink for the sum of other components of surface energy exchange when the equilibrium surface temperature solution was equal to or greater than 0°C. Melting was terminated when the equilibrium temperature descended below 0°C and the soil heat flux was computed using the thermal conductivity of the glacier assuming the depth of the freezing plane to be controlled by the freezing of pore water (assumed to be 20% by volume). The results were consistent with the author's observations of night crust thickness and pre-dawn snow surface radiant temperatures.

In short, the summer temperate glacier is considered as a plane having a slope and exposure which can be statistically described by fitting a plane to the cirque ridge rim (Lee, 1963). In the model the mean thermal and optical properties of ice and snow are weighted by the ice fraction, a dimensionless number indicating the fraction of the glacier surface area which is ice covered. In this case, ice and snow albedo values were set at .4 and .7 and the thermal conductivities at .220 and .0025 W $m^{-1}deg^{-1}$. The radiation temperature of the sky hemisphere was arbitrarily set at 25 degrees below the mean diurnal air temperature. The simulator so arranged calculated a synthetic energy transfer, surface thermal regime and a total diurnal melt generation estimate. Simulation experiments in climatology have the great advantage of liberating the investigator from the restrictions of an historical data base to perform sensitivity analyses on individual boundary conditions effects within the simulated system. Thus, insights into the nature of the operating system which are obscured by the simultaneous variation of boundary conditions in field data are made possible.

The Test Simulation

The possession of a simulator for temperate glacier runoff in addition to the general surface climate simulator will permit an exploration of the general relationship between degree-day units generated over an optically dark land surface and glacier runoff. The glacier is assumed to be located at 40°N and to be sloping 15° to the northeast. The aerodynamic roughness length at the glacier surface was fixed at .1m and the ice fraction was set to .5 yielding an integrated surface albedo of .55. The meteorological conditions assumed uniform over both surfaces were a wind velocity of $1\ m\ s^{-1}$ and 80% relative humidity. The land area was assumed to be a horizontal area with an albedo of .18, a thermal conductivity of $.26 Wm^{-1} deg^{-1}$ and a volumetric heat capacity of $2.1 KJ\ Kg^{-1} deg^{-1}$. The land area is further assumed to have local obstructions approximately 4m high that present a vertical "seen" silhouette to beam radiation and the wind of .1 the horizontal area. The surface is assumed to be representative of upland tundra or mountain meadow conditions where a weather screen might be placed by a glaciologist during a study of a nearby glacier. A variable surface climate was generated over both surfaces by varying the air temperature in five cases and the solar declination in five additional cases. The results are listed in Table One. Regression equations may be fitted to these data. The data set was split into two groups based on the source of climatological variation. There were five cases in the temperature set and five cases in the solar radiation set. The results of this analysis are reported in Table Two.

These results clearly indicate the degrading effect of temperature variation on the correlation coefficient of the pooled data set. Here we have joined Andrews et al (1971) as we are discussing parameters derived in a system built on considerations which are violently non-isomorphic with the reality of snow hydrology! Presumably similar degrading effects could be produced by variation in any of the geographic or meteorological boundary conditions required by the simulator. Note that the simulator reported melt at -10° under a high solar radiation flux at summer solstice in agreement with field observation.

TABLE ONE

Simulation Results

Solar Declination (°N)	Mean Diurnal Air Temperature (°C)	Degree-Days Over Land	Daily Melt Generation at Glacier (cm H_2O)
23	10	16.11	3.77
23	5	12.09	3.40
23	0	11.09	2.84
23	-5	10.94	2.25
23	-10	8.87	1.67
20	10	15.75	3.54
17	10	15.38	3.31
14	10	15.01	3.07
11	10	14.60	2.84
7	10	14.06	2.53

TABLE TWO

Regression Reporting

Data Set	Correlation Coefficient	Significance*
POOLED	.793	$.62X10^{-2}$
T-SET	.900	$.39X10^{-1}$
T-SET	1.000	$.10X10^{-4}$

*probability of no-correlation and zero-slope

Conclusions

1) Due to the requirements of randomness, independence and homogeneity Andrews et al (1971) discarded the "baby with the bathwater" when they chose to treat hydrometeorological data within a non-isomorphic analytical system (parametric statistics) and then used the same analytical system (which was demonstrated to be non-isomorphic) for a discussion of the synthetic "high correlation." In effect, this is asking the baby why he was in the bathwater after the fact.

2) As the major design consideration in climate simulator construction is to make the product isomorphic with reality, the application of these methods should be valuable in designing and operating field runoff prediction systems and in the analysis of the environment of regression

residuals in old data. Interested individuals may obtain a Fortran IV listing of the Glacier-Runoff Simulator from the author.

REFERENCES

Andrews, J.T., B.D. Fahey and D. Alford, 1971: "Note on Correlation Coefficients Derived from Cumulative Distributions with Reference to Glaciological Studies", Journal of Glaciology, Vol. 10, No. 58.

Bruce, J.P. and R.H. Clark, 1966: Introduction to Hydrometeorology, Pergamon Press, Ltd., London.

Lee, R., 1963: Evaluation of Solar Beam Irradiation as a Climatic Parameter of Mountain Watersheds, Hydrology Papers, Colorado State University, No. 2.

Outcalt, Samuel I., 1971: "A Numerical Surface Climate Simulator", Geographical Analysis, (in press).

Outcalt, Samuel I. and D.D. MacPhail, 1965: "A Survey of Neoglaciation in the Colorado Front Range", University of Colorado Studies Series in Earth Sciences, No. 4.

Yong, R.N. and B.P. Warkentin, 1966: Introduction to Soil Behavior, The MacMillan Co., New York.

NEWS AND COMMENTS

The 1973 meeting of Friends of Climatology was held in Montreal March 30 and 31. The afternoon of Friday, March 30, was devoted to a discussion on "Climate and the City", which was held at McGill University. Morris Charney, urbanist and architect, was chairman for a session with three speakers: Conrad East, Directeur du Centre d'Etudes Ecologiques à l'University du Québec à Montréal, gave a talk on "An Evaluation of Air Pollution in Montreal"; Daniel LaFleur, Adjoint Administratif au Directeur de la Région Richelieu de l'Hydro-Québec, spoke on "Une Etude du Confort Humain dans la ville de Montréal"; and Don Boyd, of the Division of Building Research, NRC, Ottawa, concluded the session with a talk on "Climate and Building in Northern Canada". Each talk was followed by discussion from the floor which, in a bi-lingual atmosphere, revealed the considerable and lively interest of the 70-80 people making up the audience.

A cocktail party and dinner was held in the McGill Faculty Club on Friday evening. Liquid refreshment and background music by an accomplished accordionist (Vernon Bergstrom) provided a suitable introduction to a lively and spirited discussion on "Opportunities for Climatology and Climatologists". This was a free-for-all and resulted in some revealing remarks on the need for more publicity for the work being done by climatologists, criticism of existing methods and facilities for taking observations, questions on the value and direction being taken in urban climatology, and the problems of making openings in various fields - notably agriculture - where climatologists might be thought to be able to provide a useful contribution.

On Saturday morning (March 31) the meeting continued at Sir George Williams University where the theme for discussion was "Climate and Industrial Development". Peter Barry, Senior Research Scientist, Atomic Energy, Chalk River, spoke on "Pollution Climatology", followed by Norman Powe of the Dorval Meteorological Office (Atmospheric Environment Service) on "The Relevance of Climatological Data", and L.D. Tufts, of C.P. Rail, on "Climate and the Railroad Industry". The session ended with a presentation by John Lewis (University of Maryland) and Sam Outcalt (University of Michigan) who explained an interesting system of urban climatic modelling using infra-red photography and ground control data.

During the course of a short "business session" included on Saturday morning it was suggested that a register of climatologists working in North America should be prepared, and that the Friends might also consider organising workshops to discuss recent methodological developments and research procedures. B.J. Garnier undertook to develop both these suggestions further in conjunction with the persons who introduced them and spoke in support of them at the meeting. The meeting also accepted with acclamation Wolfgang Baier's offer to organise the 1974 meeting in Ottawa sometime between January and the end of March.

The Meeting ended with a vote of thanks to those who hade made the arrangements in Montreal proposed by Morley Thomas.

Some Recent Publications of interest to readers of the BULLETIN are:

(a) "Weather Systems and Precipitation Characteristics over the Arctic Archipelago in the Summer of 1968" by Stephen Fogarasi, Scientific Series No. 16, Inland Waters Directorate, Water Resources Branch, Ottawa, 1972.

(b) "Winter Clothing Requirements for Canada" by Andris Auliciems, Christopher R. de Freitas, and F. Kenneth Hare, Climatological Studies No. 22, Atmospheric Environment Service, Toronto, 1973.

(c) "A Guide to the Records and Microfilm of McGill Observatories and some Records of other Weather Observers" from 1798-, University Archives, McGill University, Montreal, December 1972.

Tom Dunne, who has been Assistant Professor of Hydrology at McGill University for the past four years, is leaving to take up a position in the Geology Department of the University of Washington in Seattle. His place at McGill will be taken by John Drake now completing his doctoral degree at MacMaster University.

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No. 4 Temperature Variability and Synoptic Cold Fronts in the Winter Climate of Mexico, by J. B. Hill, 71 pp., February 1969, price \$7.50.

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