SNOWMELT COMPUTATIONS FOR A HIGH ARCTIC SITE

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ABSTRACT

The energy balance approach was used to calculate the snowmelt at a site near Resolute Bay, Northwest Territories (75°N, 95°W). Snowmelt energy was partitioned into net radiation, sensible heat, latent heat and rain melt components. Net radiation and sensible heat fluxes accounted for approximately 60 and 40 percent of the incoming energy while snowmelt and evaporation consumed 80 and 20 percent of the energy respectively. This study demonstrates that unless the snow cover is extremely patchy, the energy balance approach adequately estimates the rate of snowmelt at a High Arctic site.

INTRODUCTION

In the non-glacierized basins of the High Arctic, meltwater is responsible for the bulk of the annual runoff (Marsh 1978) . Over 80 percent of the total runoff occurs within a period of three weeks. Despite the hydrologic significance of snowmelt floods in most High Arctic basins, only a limited number of studies has been carried out in the Arctic (Weller and Holmgren 1974, Ohmura 1972, Müller and Keeler 1969). The present study was therefore carried out to enable an understanding of the energy supplies and losses to an Arctic snowpack and to determine the feasibility of a surface energy balance approach in the prediction of melt in an Arctic environment.

SURFACE ENERGY BALANCE

At the snowpack surface, the energy balance is given by

$$Q_M = Q^* + Q_H + Q_E + Q_P + Q_G \qquad (1)$$

where Q_M is the energy available for melting, Q^*, Q_H and Q_E are the fluxes of net radiation, sensible and latent heat. The energy added by precipitation and the ground heat flux are represented by Q_P and Q_G respectively.

The terms on the right hand side of equation (1) are to be evaluated or measured in order to obtain Q_M. The net radition flux can be readily measured with a radiometer. The evaluation of Q_H and Q_E requires the measurement of temperature, humidity and windspeed at two or more levels above the surface. However, during the melt period, the snow surface is nearly 0°C and the surface humidity is close to 100 percent. This permits the use of the bulk transfer approach to calculate the fluxes of sensible and latent heat.

$$Q_H = \rho_a C_a D_H u (\theta_z - \theta_s) \qquad (2)$$

$$Q_E = \rho_a \lambda (\varepsilon/P) D_E u (e_z - e_s) \qquad (3)$$

Here,

θ_s and θ_z are temperatures at the snow surface and at a height z metres above the surface (K)

e_s and e_z are the vapour pressures at the snow surface and at a height z metres above the surface (Pa)

ρ_a is the air density (kg m^{-3})

C_a is the heat capacity of air at a constant pressure ($Jkg^{-1}K^{-1}$)

λ is the latent heat of vapourization (Jkg^{-1})

ε is the ratio of the molecular weights of water and air (dimensionless)

P is the atmospheric pressure (Pa)

u is the wind speed at height z (ms^{-1})

D_H and D_E are dimensionless drag coefficients

It is assumed that $D_H = D_E = D$

$$D = K^2/[\ln (z/z_o)]^2 \tag{4}$$

where K is the von Kármán's constant (dimensionless)

The variable z_o is a measure of the surface roughness (in m) and can be estimated from wind speed measurements of two levels, z_2 and z_1 if a logarithmic wind profile is assumed

$$z_o = \exp[(u_2 \ln z_1 - u_1 \ln z_2)/(u_2 - u_1)] \tag{5}$$

The bulk transfer equations are valid only for neutral atmospheric conditions. To correct for the frequent occurrence of temperature inversions or stable conditions over snow, the drag coefficient can be modified (Price et al. 1976) using the Richardson number, Ri,

$$Ri = gz(\theta_z - \theta_s)/\theta_z(u_z - u_s)^2 \tag{6}$$

where g is the acceleration due to gravity (ms^{-2}). For stable conditions ($Ri > 0$), and the drag coefficient is

$$D_s = D/(1 + \sigma Ri) \tag{7}$$

where σ is a coefficient with a value of 10. Unstable conditions ($Ri < 0$) are less common, but the drag coefficient can be modified by

$$D_u = D(1 - \sigma Ri) \tag{8}$$

The other two terms of the energy balance equation (1) can be evaluated by

$$Q_p = C_w \rho_w R(\theta_R - \theta_s) \tag{9}$$

where C_w is the specific heat of water ($Jkg^{-1}K^{-1}$) and ρ_w is the density of water (kg m^{-3}), θ_R is the temperature of the rain which can be approximated by the wet bulb temperature and R is the rainfall intensity (ms^{-1}).

$$Q_G = -\kappa(d\theta/dz) \tag{10}$$

where κ is the coefficient of thermal conductivity ($Js^{-1}m^{-1}K^{-1}$) and $d\theta/dz$ is the temperature gradient in the ground.

In the High Arctic, rainfall is usually small and the contribution of Q_p becomes negligible. The ground heat flux is also small and in the computation of the snow surface energy balance, this term can be considered as zero.

EXPERIMENTAL SITE AND SNOW CONDITIONS

The study site was located approximately 3 km northeast of Resolute Bay (75°N,95°W) on Cornwallis Island, Northwest Territories (Fig. 1). The site elevation was 76 m above sea level and was surrounded by gently rolling topography with hills rising to a height of 170 m.

This study was carried out in June 1977. Prior to the melt season, there was 95 mm of snow (in water equivalent units) in the drainage basin where the site is situated. Woo and Marsh (1977a) have commented upon the uneven distribution of snow cover, being deeper in the valleys but very thin on the ridges and hill tops. The winter of 1976-77 received light snowfall. The amount was less than 70 percent of the total snowfall of the previous winter (Woo and Marsh 1977b). Thus at the end of May, many ridge tops and exposed surfaces were bare and large expanses of flat areas had a thin veneer of snow.

Immediately before the melt period the temperature of the snowpack averaged -10°C, but the snowpack began to ripen in early June, becoming isothermal at 0°C by June 12th (Fig. 2). By then the snowpack had a basal layer of ice together with numerous ice lenses within the pack.

DATA COLLECTION

In June 1977 a site was set up to measure temperature, humidity and windspeed at a height of 1 m above the snow surface. Air temperature and humidity were recorded by a Lambrecht thermohygrograph, and the recorder readings were frequently checked by a mercury thermometer and by an Asman psychrometer. Net radiation over snow was measured by a Swisstecco net radiometer whose signals were recorded by a Rustrak recorder after appropriate amplification. Wind speed was measured by a Munro cup anemometer, but dial readings were taken only during the day. To correct for this deficiency, the site wind speed was compared against the 10 m wind speed recorded by the Atmospheric Environment Service at Resolute Bay, and an empirical relationship thus obtained was used to provide hourly wind speed for the site. The weather station also provided data on air pressure and rainfall. z_o was obtained experimentally and the average value was 0.003 m.

The snow surface temperature was observed to fall below 0°C on only two occasions, but in neither case did it fall below -0.2°C. Although snow surface temperature was not measured between 2200 h and 0700 h, it is reasonable to assume that the temperature was 0°C throughout the period.

Using the data thus obtained, the snow surface energy balance was evaluated hourly to provide estimates of Q_M. The calculated values of snowmelt were then compared with the ablation measurements made at the study site. Snow ablation rate (M) was determined by

$$M = (h_t - h_{t+\Delta t})/\rho_s \qquad (11)$$

where h_t and $h_{t+\Delta t}$ are the average heights of the snow surface above an arbitrary datum at times t and $t+\Delta t$. These heights were obtained by averaging the distance of approximately ten points on the snow surface below a taut wire whose ends were firmly held in position by two rods frozen into the ground. ρ_s is the snow density obtained during the time period Δt using 250 cm^3 samples of surface snow cores.

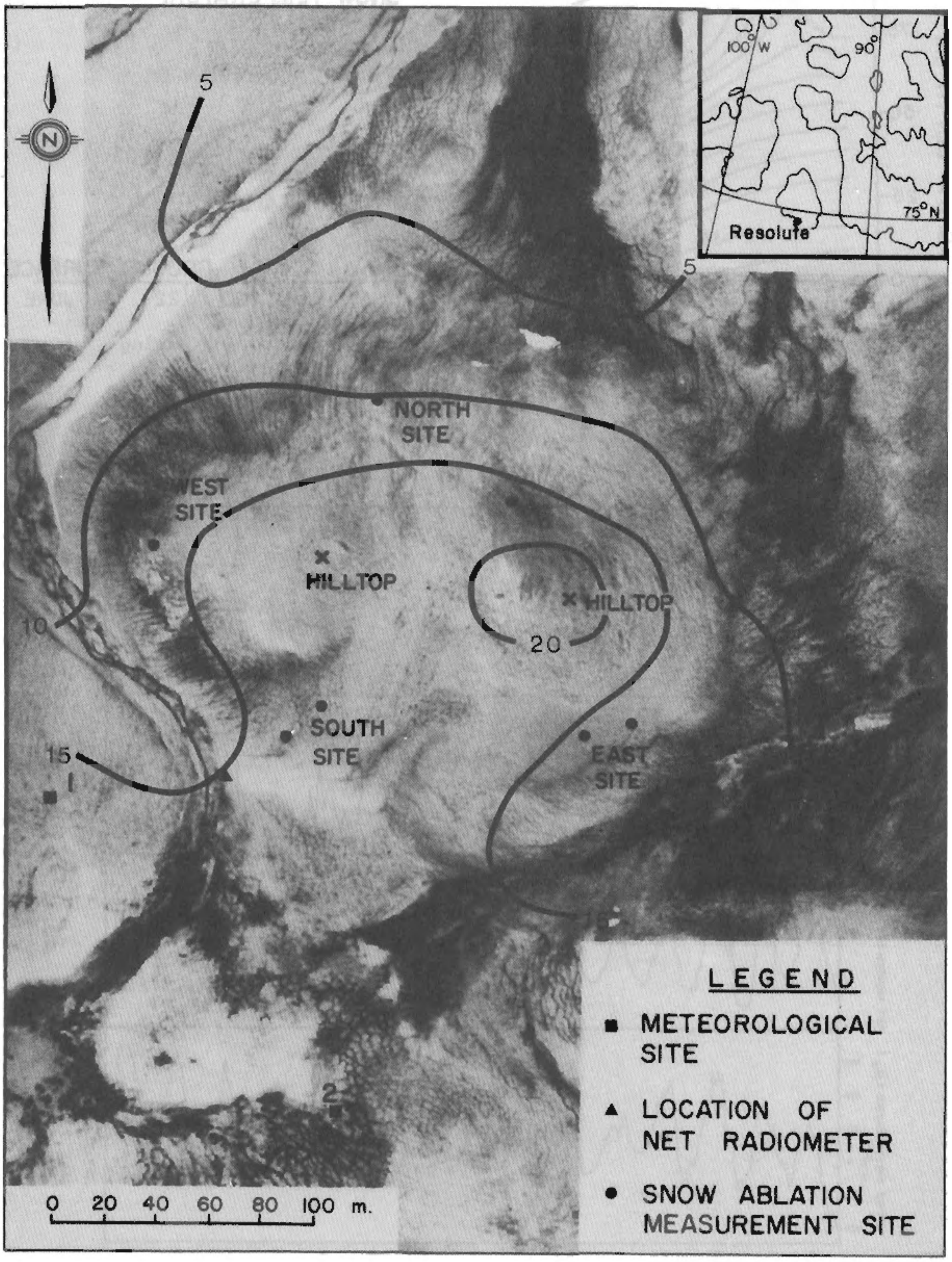

Fig. 1 Location of the experimental site, 3 km northeast of Resolute. Contours are at 5 m intervals above an arbitrary datum.

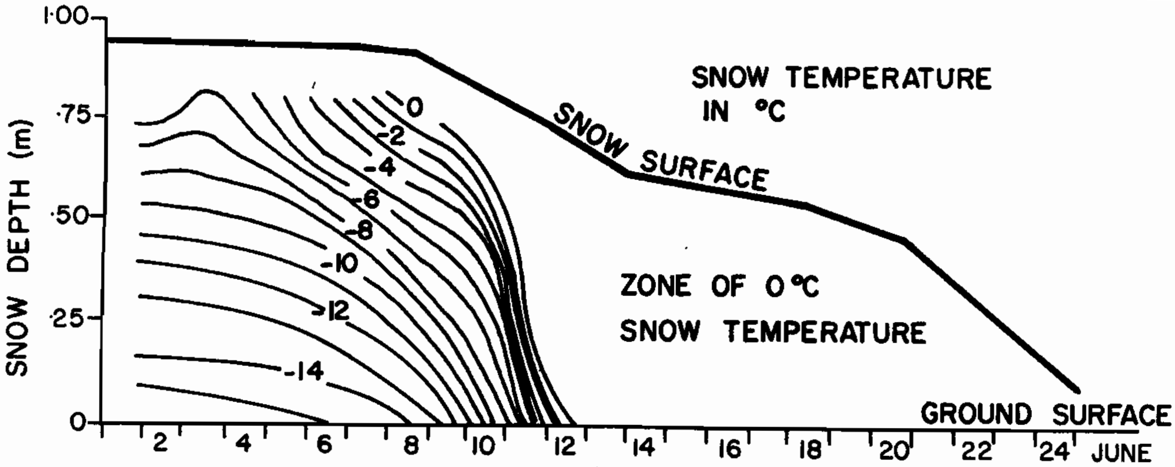

Fig. 2 Changes in the snowpack temperature before and during the study period.

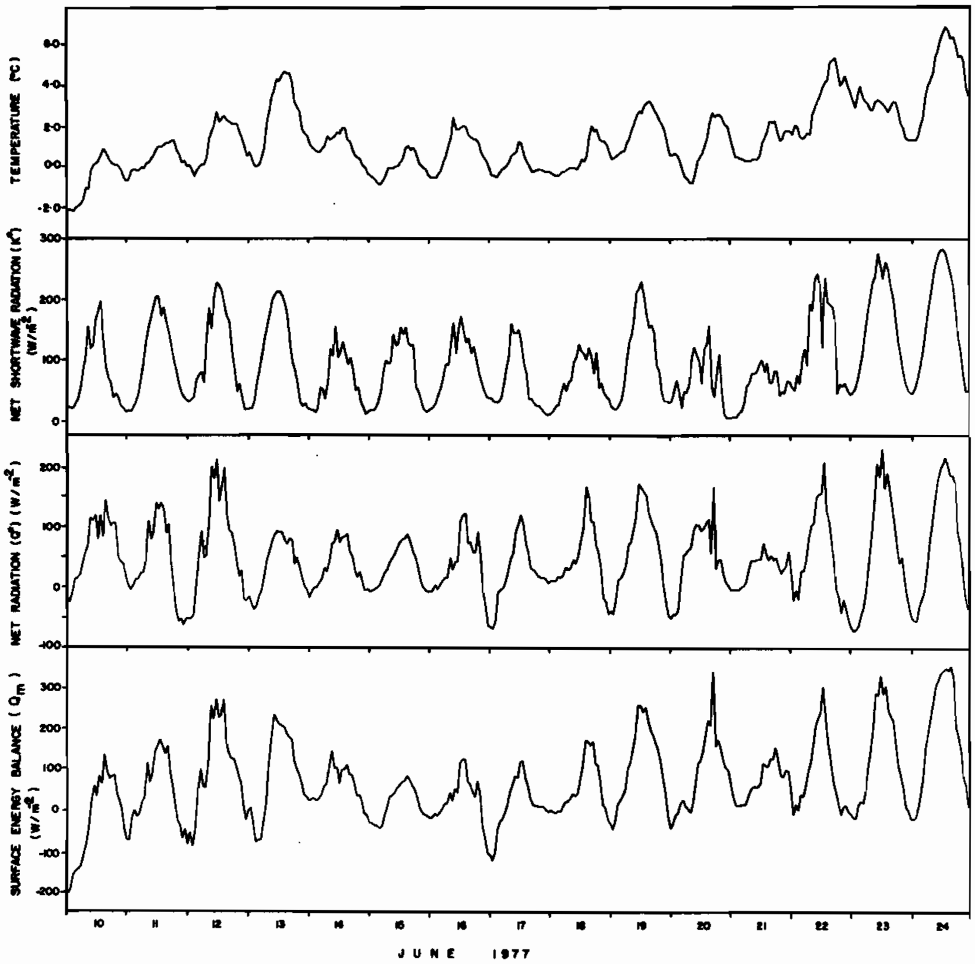

Fig. 3 Variation of air temperature and radiation during the melt period.

RESULTS

Before June 10th, most of the available energy was used to ripen the snowpack. On June 10th, the air temperature rose above 0°C and remained so except for at most a few hours each day (Fig. 3). On cloudy days, net radiation remained above zero for 24 hours. On clear days, negative net radiation occurred during the early morning hours, but continuous daylight prevented large negative values (Fig. 3).

The melting of a High Arctic snowpack was rapid. Many areas became snowfree soon after snowmelt began. A general increase in snowfree areas near the study site is shown in Figure 4 which covers a period of 21 days.

Figure 3 also plots the hourly surface energy balance (Q_M) which closely follows the pattern of net radiation. The hourly melt values are accumulated and superimposed on the cumulative measured ablation curve (Fig. 5). On the whole, the computed and measured values correspond closely, indicating the applicability of the energy balance approach to estimate snowmelt. Several slight discrepancies are noticeable. One of them occurred during the first three days of measurement when the snowpack was not completely ripe. Hence, some of the ablation measurements probably included the settling of the snowpack. This is evidenced by an increase in snowpack density from 350 kg m^{-3} in the premelt period to 450 kg m^{-3} (ignoring ice lenses) shortly after ripening. Another disagreement between the measured and the calculated data occurred on June 17th and 18th when overcast and foggy conditions prevailed. Light drizzle fell but the intensity was not measurable. This possibly resulted in an underestimation of rain melt. A third period with discrepancies was when extensive areas of bare ground produced an advection of sensible heat flux to the snow patches. This was very prominent after June 22nd when the snow patch was so reduced in size that the meteorological measurements were probably not representative of the boundary layer.

On a daily time scale, Figure 6 shows a comparison of calculated snowmelt against observed ablation. The points where calculated values underestimated actual ablation occurred during the early stages of melt, while the points of over-estimation occurred at the end of the main melt period.

Within each day, the hourly distribution of energy fluxes vary. Figure 7 shows the energy distribution for two typical days. On this day (June 15th), both the net radiation and sensible heat fluxes were small. High humidity and near freezing temperatures resulted in low vapour pressure gradients and a low latent heat flux. Although the surface energy balance was small, the lack of any negative net radiation produced almost continuous melt throughout the 24 hours period.

On a clear day with scattered clouds (such as June 12th), the net radiation component was larger during the day. The exchange of sensible heat peaked after net radiation as the air became warmer in the afternoon. Both sensible and latent heat fluxes produced prominent diurnal cycles which tended to mirror each other. Similar daily cycles have been observed over prairie snowpacks by Male and Gray (1975).

The various components of energy flux were partitioned on a daily basis for the entire study period. The ground heat is assumed negligible and rain melt was zero or almost zero. Then, the heat flux to the snow is the sum of the net radiation, sensible heat and latent heat components. Figure 8 shows that on overcast or foggy days between June 14th and 17th, energy available for melting was reduced. Latent heat flux was predominantly an energy loss from the snow surface, but its magnitude decreased after mid-June. The fluxes of net radiation and sensible heat contributed most of the snowmelt energy. For the entire period of measurement, net radiation provided 58 percent of the total incoming energy while sensible heat exchange accounted for 42 percent of the energy input to the surface. Eighty percent of the energy received was used to melt the snow and the remainder was consumed in evaporation.

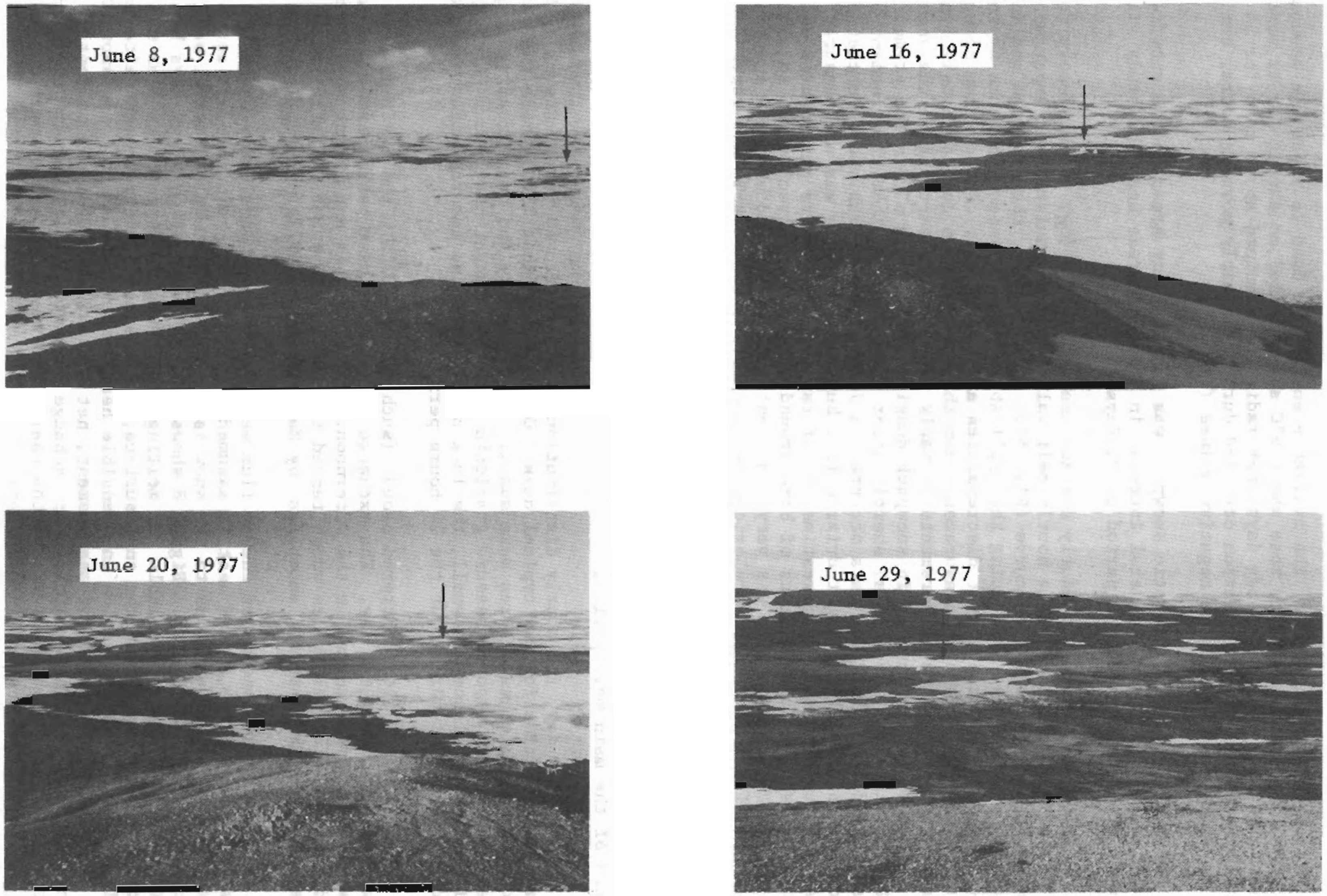

Fig. 4 Photographs showing the spatial variation of the snowcover in the vicinity of the experimental site. The arrow points to meteorological site 1.

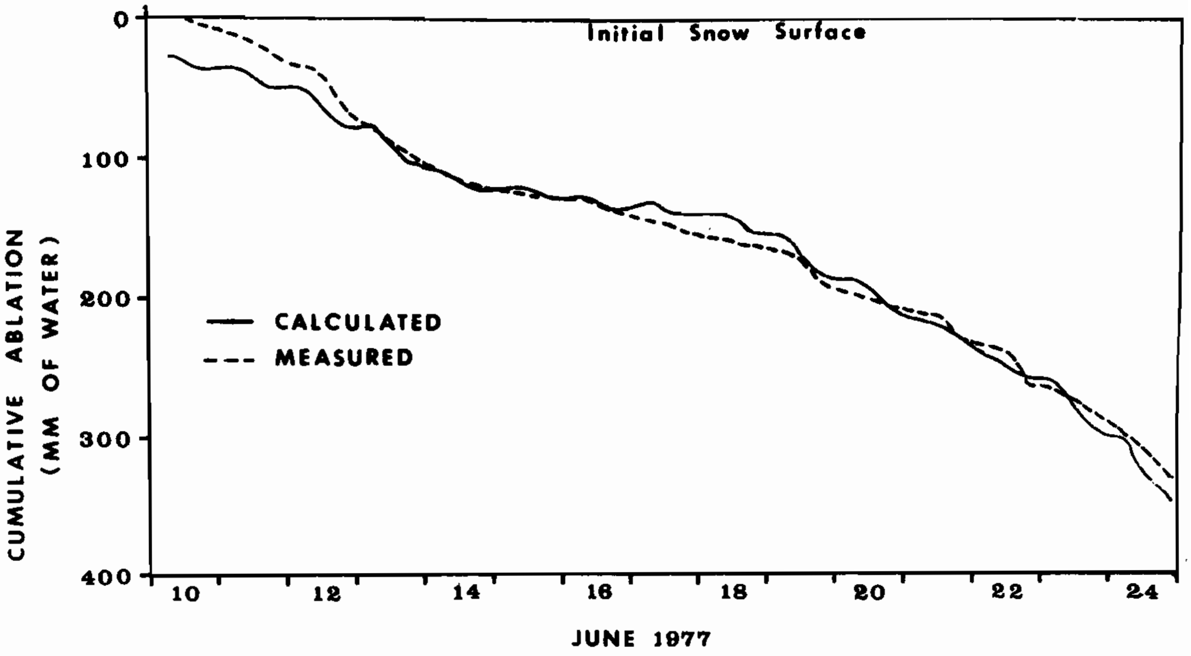

Fig. 5 Comparison of cumulative calculated snowmelt and cumulative measured ablation.

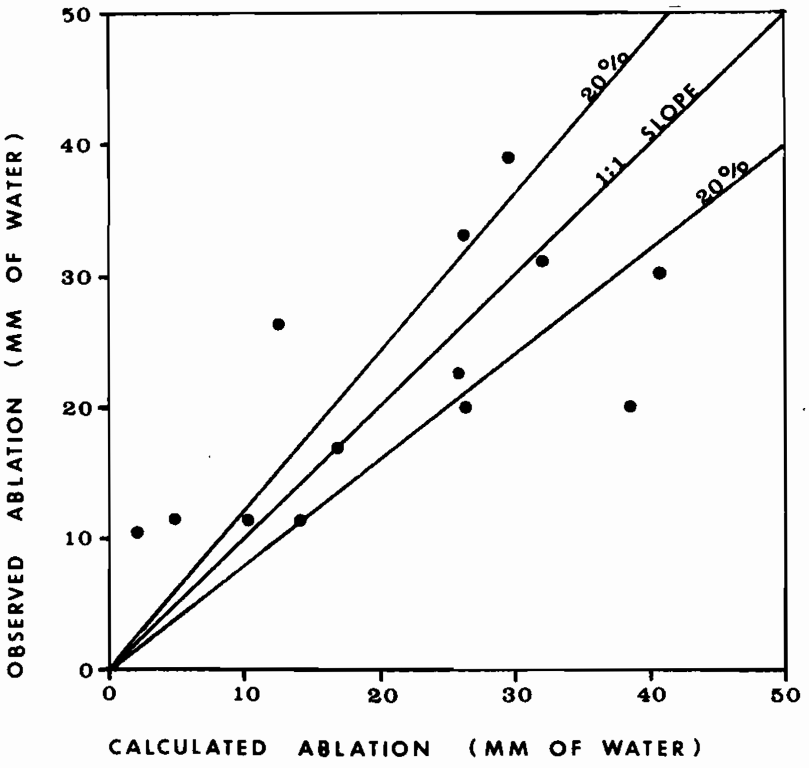

Fig. 6 Comparison of daily calculated snowmelt and daily observed ablation.

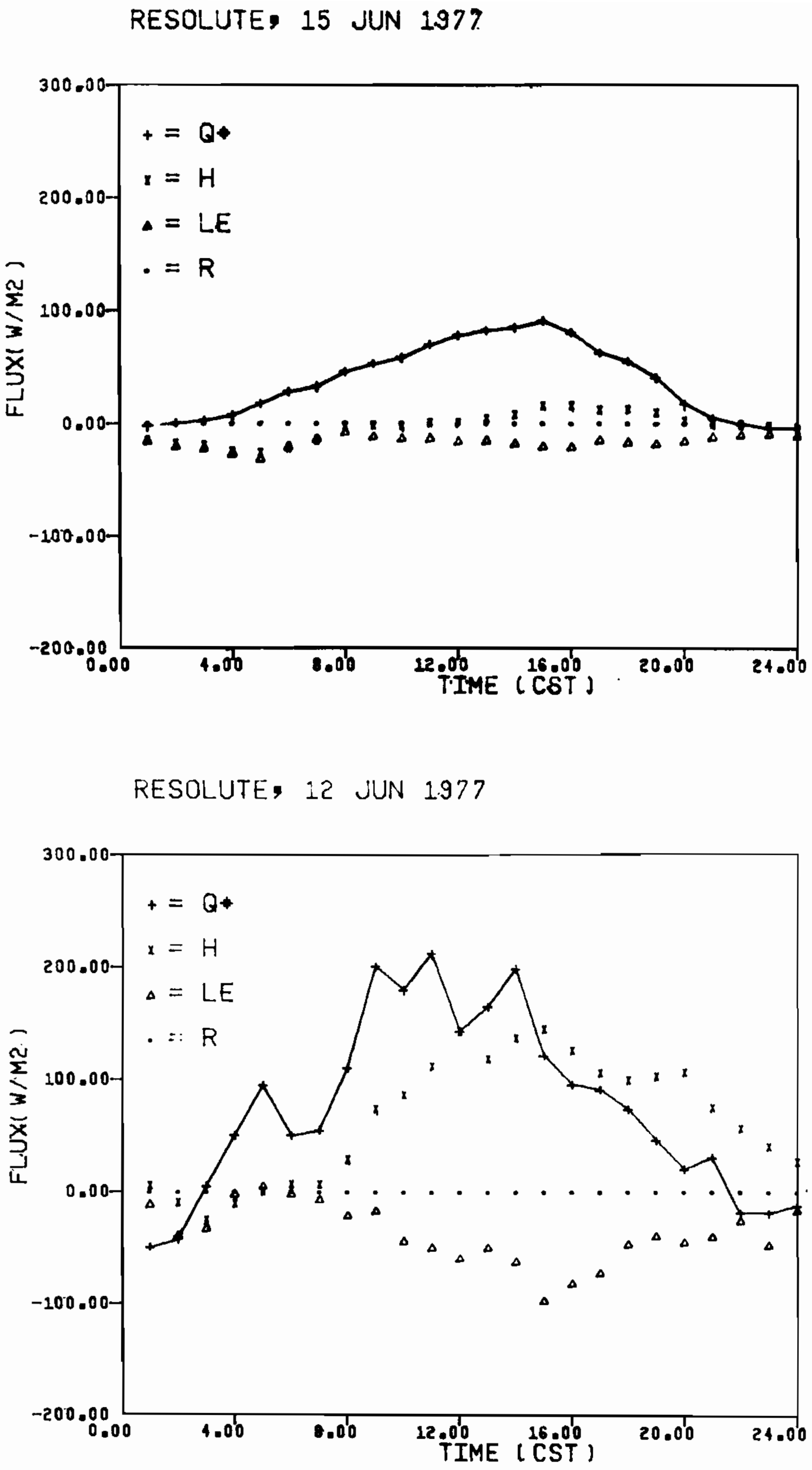

Fig. 7 Diurnal variation of the surface energy balance components during an overcast day (above) and a clear day (below). Q^* is the net radiation, H is the sensible heat flux, LE is the latent heat flux and R is the energy gained from precipitation.

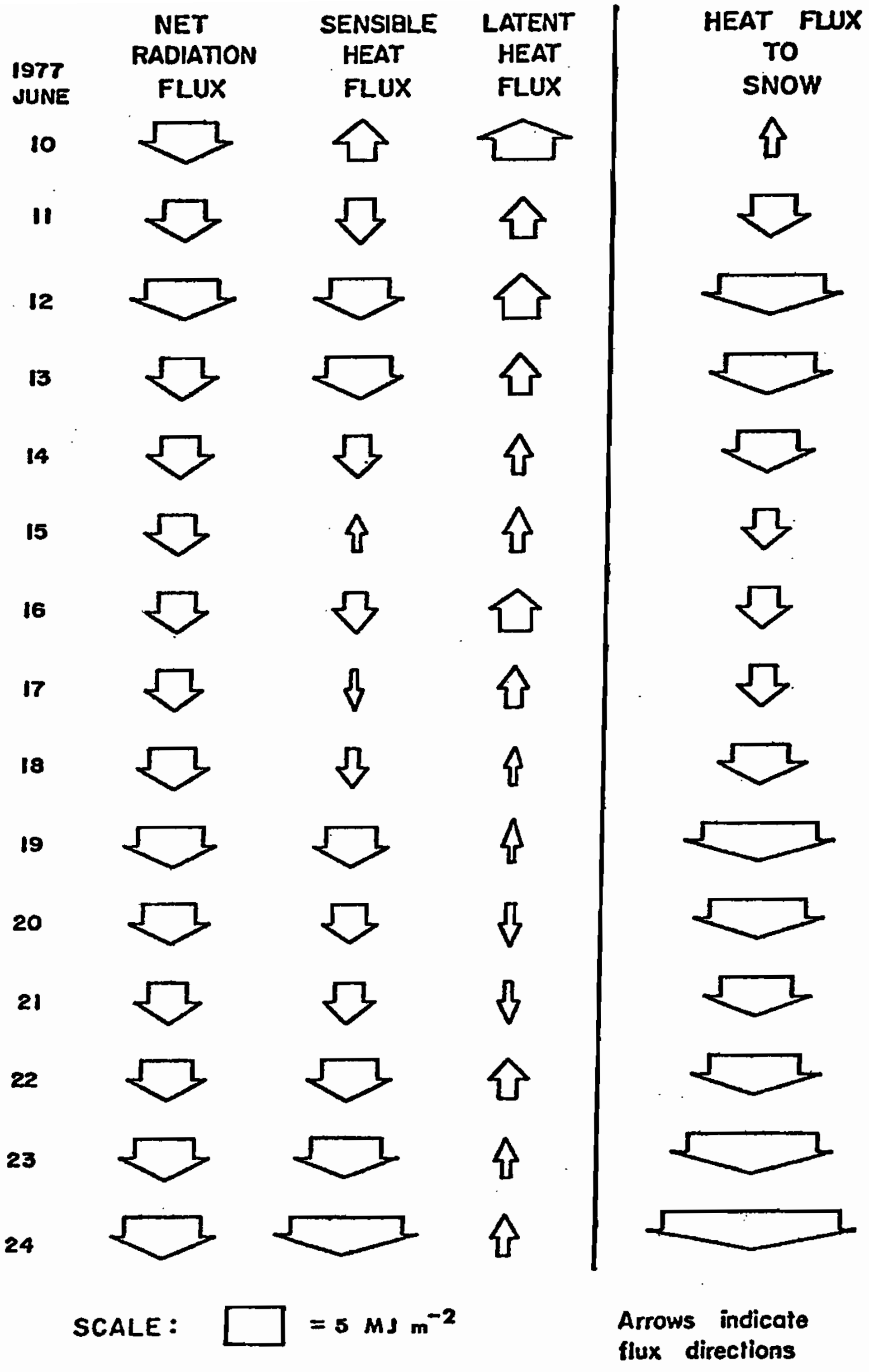

Fig. 8 Daily variation of the energy balance components. The width of the arrow indicates the relative magnitude of energy fluxes.

CONCLUSIONS

This study demonstrates that snowmelt at a High Arctic site can be successfully computed by the surface energy balance. Latent heat flux to the snowpack was usually negative, but the magnitudes of net radiation and sensible heat flux were able to sustain both evaporation and snowmelt. During clear days, both net radiation and sensible heat flux increased, resulting in an increase in melt. On these days, conspicuous diurnal cycles in the fluxes of net radiation, sensible heat and latent heat were observed. On overcast days, net radiation became the dominant energy input while the fluxes of sensible and latent heat were small.

The energy balance approach becomes inaccurate when a continuous snow cover disintegrates into isolated snow patches. This is due to an advective flux of sensible heat from the bare ground nearby. Thus, if the snowcover remains continuous for much of the melt period, the net radiation component should account for most of the incoming energy. Otherwise, an incomplete snowcover will increase the importance of the incoming flux of sensible heat.

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