SNOWMELT AT THE HIGH ARCTIC SITES, RESOLUTE, N.W.T., CANADA

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ABSTRACT

To understand the snowmelt processes in the high Arctic, field work was conducted during late May to mid June of 1994 at two sites in the study area near Resolute, Cornwallis Island. One site had relatively clean snow and the other was covered by windblown dust from the road and runway nearby. Microclimate elements (air temperature, precipitation, humidity, wind speed, radiation) and snow temperature and daily snow melt were recorded by automatic instruments at both sites. Snow samples were collected at fixed locations on selected dates for dust content analysis. Comparison of the microclimate was made between the sites and significant difference of dust content, of snow surface albedo and, hence, of net radiation was found. As result of the radiation difference, the dirty snow melted earlier and the snow cover disappeared about 10 days before the clean snow.

In the calculation of energy balance, snow surface roughness length was estimated by wind profile method using wind observations at 3 levels at the site. The comparison of snowmelt computed from the energy balance was made to the observed snowmelt in the filed. A Good agreement has been obtained at two sites with different snow conditions. This indicates that the energy balance approach is a good short-term predictor of snowmelt in the high Arctic.

INTRODUCTION

Spring in the polar regions is a very important transition period with rapid changes in the radiation balance and large-scale atmospheric circulation. In this period, snowcover melts, surface albedo changes drastically and streamflow initiates. Many studies have reported field observation and energy balance calculation of snowmelt in Arctic (Price and Dunne, 1976; Dunne et al., 1976; Heron and Woo, 1978; Ohmura, 1982; Woo et al., 1982; Hardy, 1993).

The present paper reports the results of the application of the energy balance method to the simulation of daily snowmelt at two sites in high Arctic. These sites are typical of areas with windblown snow and areas near human settlements where the snow is dirty. Field observation of snow condition, albedo, ablation rates were compared with energy balance calculations of snowmelt. Our results should be useful for climate model and hydrological study in the polar regions.

STUDY AREA AND FIELD METHODS

This study was conducted in Resolute, Cornwallis Island, N.W.T., Canada. The Climate of Resolute is characterised by relatively low precipitation (annual total 131mm) and extremely cold temperature (annual mean -16.6°C), with very strong wind (annual mean 5.9m/s). The mean annual total snowfall was reported about 84cm at Resolute weather station. Typically snowpack formation starts in early September. The snowpack remains cold until early May when it reaches its maximum depth of about 30-60cm. Blowing snow occurs frequently, leading to the extreme uneven snow distribution over various terrain types (Woo and Marsh, 1978; Woo et al., 1983; Woo et al., 1995).

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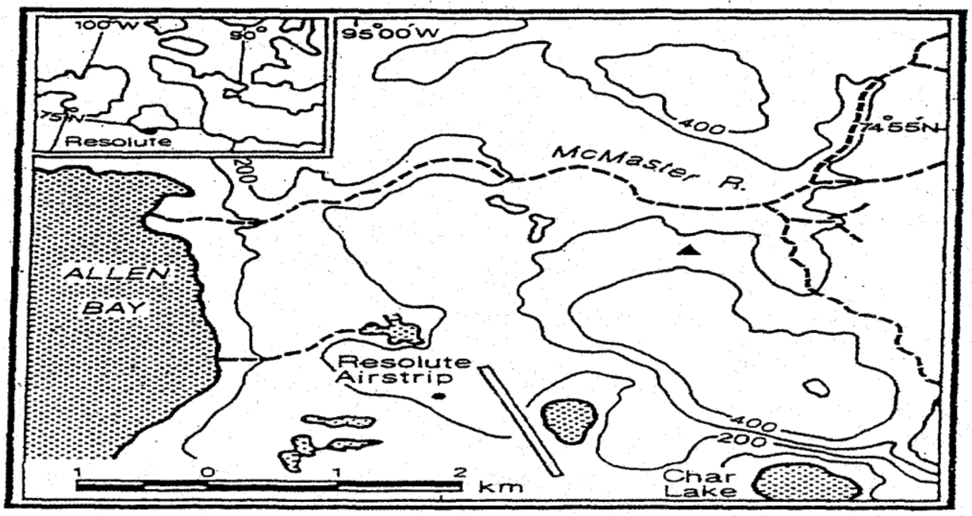

• runway AWS ♦ McMaster basin AWS --- stream 100' contour interval

Fig. 1. Location and topography of the study sites, Resolute Bay, Cornwallis Island, N.W.T.

Two sites about 3km apart were chosen for the study: one near the runway of the airport and the other in McMaster river basin (Fig.1). Automatic weather stations (AWS) were operated at the sites. The variables measured at the sites included global radiation ($K\downarrow$), reflective radiation ($K\uparrow$), net radiation (Q^*), air temperature, humidity, wind speed at 3 levels and precipitation. Internal snowpack temperatures were measured within 20m of each station by thermocouple (10-20 cm vertical spacing) and snow depths were observed both manually at snow stakes and by an ultrasonic depth gauge (UDG10). Data at both sites were measured at 1min intervals with identical sensors and averaged over 60min and recorded hourly with a Campbell Scientific CR-21x or a CR-10 dataloger. Snow boards and a rain gauge were used for precipitation measurement. Visual observations comprised the cloudiness, snow accumulation, and the type of precipitation.

Spatial distribution of snow depth and snow density at the two sites was determined prior to melt by a snow survey around the AWS. Since variability of snow depth generally exceeds that of snow density, 35 snow depths and 5 snow density measurements were taken. Snowpack density was measured by the standard MSC snow sampler. Snow water equivalent (SWE) was calculated by multiplying mean depth and mean density. Snow pits were opened prior to the melt and on selective days during the snowmelt period, to detect the growth of the basal ice (Woo et al., 1982).

Snow survey on May 16, 1994 showed that snow depth ranged from 10 to 35cm and snow density changed from 0.295 to 0.350 at the runway site. Snowpack was deeper at the sheltered McMaster met. site, with the depth ranging from 16 to 47cm and density between 0.317 and 0.382. The snow water equivalent was determined to be 65mm at the runway site and 145mm at McMaster met. site, respectively. Snow pits were also dug on the same day, no ice layers in the pack or basal ice on the ground were found.

Woo and Dubreuil (1985) studied the effect of dust on arctic snowmelt and reported that most of the dust concentrates at the snow surface. The dust cover on the snow seldom exceeds several mm in thickness and therefore has little insulation effect. The dust often has a local source, e.g. from the bare spots on hillslopes or from the exposure surface of the road and the runway of the airport. The dust ranges in colour from light yellow through dull yellow to yellow grey. About 70 percent of the particles is sand, the rest is silt and clay. During the study period, the dust load on snow was obtained at 8:00 am on selective day by scraping off the top 20mm of an area covering 100X100 mm^2. The material collected was melted in laboratory and filtered through a pre-weighed Millipore membrane of 0.45 μm pore size. Filtering followed a field procedure described by Ostrem and Stanley (1969). The membrane was air-dried in the laboratory for 36-48 hours before bagged. The samples were brought back to Atmospheric Environment Service for re-weighing.

Snow samples, collected with the SMC standard tube, were also taken on May 16 for dust load determination, the data showed that an initial dust load of 130g/m^3 in the snowapck at the runway site versus 32g/m^3 at McMaster met. site, because the runway site was close to the dust source from the runway and the road.

Daily ablation of snowpack in precipitation-free days was measured by the snow-wire method (Heron and Woo, 1978). Snow depth change at 15 points along a 2.5m string was observed daily at 8:00am (local time) and the snow surface (top 2cm) density was measured by taking five 200cc snow samples.

RESULTS OF OBSERVATION

Air Temperature

Climate records at Resolute weather station show that monthly mean temperatures were -10.9°C for May and -0.6°C for June. During the observation period of this study in 1994 (Fig.2), a clod spell (-5°C to -20°C) was experienced and followed by a warming-up period, with the temperature rising to 0°C or slightly above the freezing point. Compared to the runway site, hourly temperature at McMaster met. site (with less exposure) was 2-5C° warmer in late May. The air temperatures were very close between the 2 sites in early June.

Good correlations of the hourly air temperature at both sites with Resolute weather station temperature records were found for the observation period; this provided confidence of accepting the observed temperature data for the energy balance calculation.

Wind Speed

The climate wind speed in May and June was 5.9-6.1m/s at Resolute, with the prevailing wind from NNW. Based on our observation, the hourly mean wind speeds at 2m height were 4.5m/s at the runway AWS and 4.2m/s at McMaster basin AWS. The highest hourly wind speed reached 13m/s and 11m/s at the 2 sites, respectively (Fig.2). Blowing snow occurred during high winds.

Comparison of the daily wind records between AES Eureka weather station and the automatic weather station at Hot Weather Creek was made by Peters and Headley (1990). They found that daily winds measured at Eureka station were not representative of other areas such as Fosheim Peninsula. In this study, the wind data collected at 2 meters above the ground at both AWS compared very well to Resolute weather station wind records (measured at 10 m height), as our AWS sites were located not far from the AES weather station.

Radiation and Snow Albedo

Fig.2 shows the hourly observation of short-wave radiation at the two sites. The mid-day incoming radiation ranged from 300 to 680W/m^2 and the mid-night values were between 20 and 80W/m^2. No increasing trend of the short-wave radiation was found in the observation period. Generally, the observations between the sites compared well during most of the hours in a day. But difference existed during noon hours when short-wave radiation reached the maximum and any difference of cloud-cover between the sites would certainly make the greatest difference on income radiation received between the sites.

Fig. 2 also shows the hourly net radiation observations at the two sites. Generally, there is a gradual increasing trend during the observation period at both sites, with the daily peak values rising from 80W/m^2 to 400w/m^2 for dirty snow and from 50W/m^2 to 150W/m^2 for clean snow. Difference in net radiation between the two sites was clearly seen through the melt period. As expected, dirty snow received more net radiation heat flux, particularly at the end of the melt period (early June for the dirty snow), due to its lower albedo caused by higher dust load and higher water content of the snowpack.

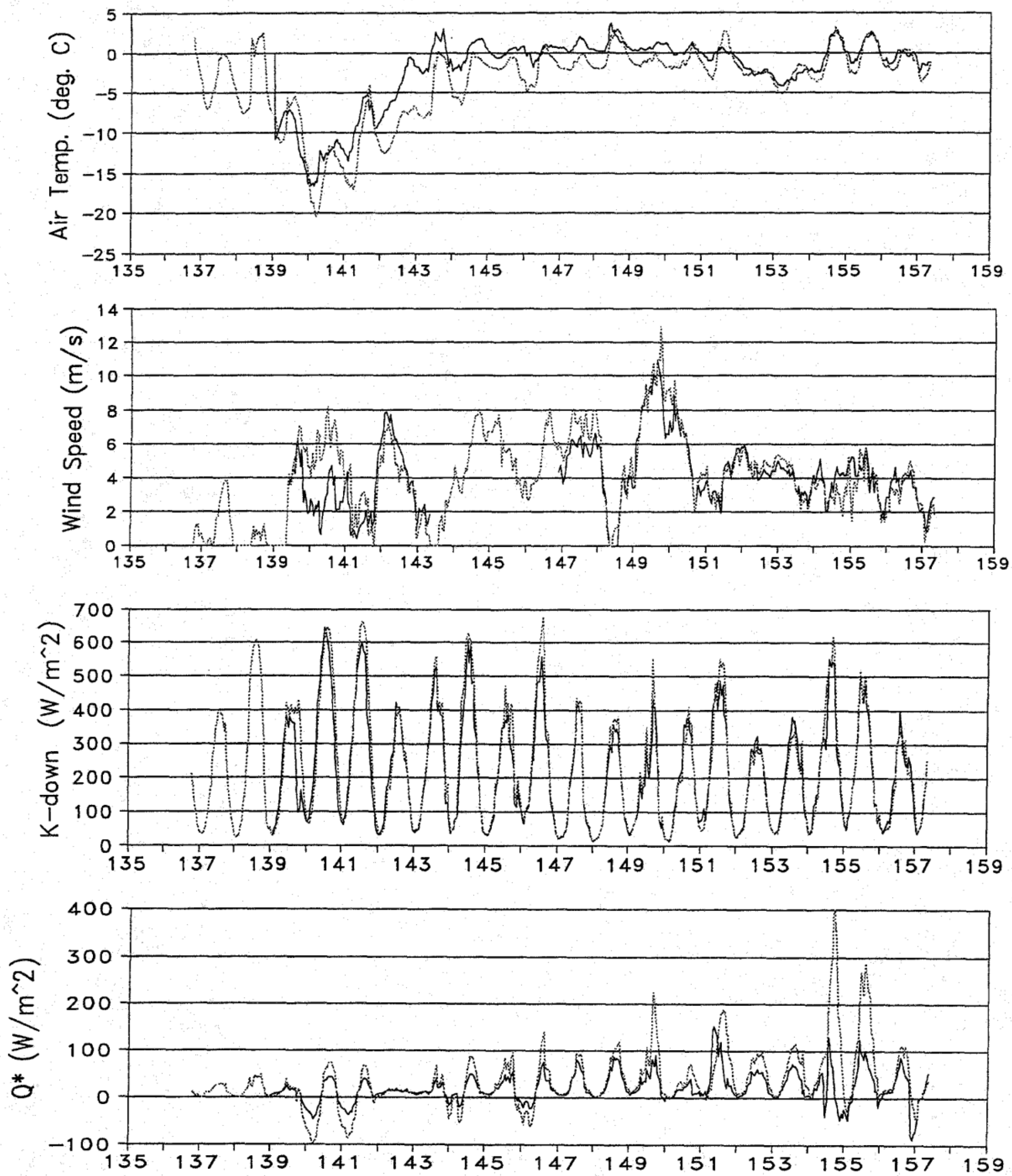

Fig. 2. Hourly observations of air temperature, wind speed at 2m, income radiation (K-down) and net radiation (Q*) at runway AWS (dotted lines) and McMaster AWS (solid lines) during May 15 (JD 135) to June 8 (JD 159), 1994.

Net radiation is the dominant energy flux for snowmelt (Male and Granger, 1981; Ohmura, 1982). The short wave radiation components often constitutes the major element of net radiation:

$$Q^* = (1-\alpha)\ K\downarrow + L^* \qquad (1)$$

where $K\downarrow$ (W/m^2) is incoming short-wave radiation and L^* (W/m^2) is net long-wave radiation. To simplify eq(1) so that L^* needs not be measured, Woo and Dubreuil (1985) developed the following empirical relation:

$$Q^* = a0 + a1\ (1-\alpha)\ K\downarrow \qquad (2)$$

where the coefficients, a0 = -20.5 and a1 = 0.69, were obtained by regression. The values were estimated from the 12-hour total radiation during the prior of 0600 to 1800 C.S.T..

Based on the radiation observations in this study, the same regression analysis was conducted for the daily total radiation data at dirty and clean snow sites. The results yielded almost identical coefficients (Fig. 3). This confirms that the empirical relation derived by Woo and Dubreuil (1985) holds during arctic snowmelt period and, thus, this approach could be applied to estimate net radiation when only short-wave income radiation data are available.

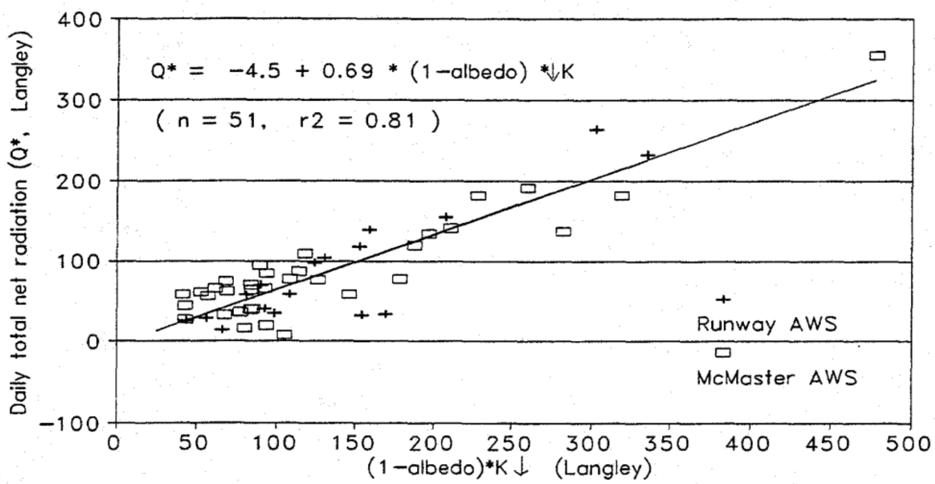

Fig. 3. Empirical relation between net radiation and short-wave radiation (daily totals in Langley).

Reflected radiation was also measured at the sites. Thus, snow albedo (reflected /incoming radiation) was obtained for each hours of a day. Fig. 4 shows the midday snow albedo during the observation period, since the sun is at the highest elevation at noon hour and albedo measurement at this time is more reliable.

Prior to the snowmelt, the midday snow albedo ranged from 80% to 90% at both sites. During the snowmelt, snow albedo dropped from 80% to 38% in 12 days at the dirty snow site. At the clean snow site, snow albedo held around 75-80% during the weak melt period before JD 164 (June 13) and then decreased sharply to 20% at the end of the melt season. The recovery of the snow albedo (back to 80-90%) was also observed after fresh snowfall events, the declining process of snow albedo continued usually on the following days (Fig.4).

Studies show many factors (such as dirt on the snow, water content, grain size and algae) affecting snow albedo (Marks and Dozier, 1992; Woo, 1984; Woo and Dubreuil, 1985; Winther, 1993; Thomas and Duaval, 1995). In this study, snow albedo was similar at dirty and clean snow sites during the initial melt period (e.g. JD 136 to 145), because the snowpack was covered by a thin layer of new snow as the result of the precipitation events (freezing rain mixed with snowfall) during May 22-25 (JD 142-145). As melt progressed, new snow melted first and the dust appeared on the melting surface of the snowpack and affected the absorption of radiation. The higher absorption of radiation by the dust caused the earlier drop of the albedo at the dirty snow site. Therefore, the difference in snow albedo between the sites became significant, particularly at the end of the melt period for the dirty snow, when water content of the snowpack was higher and more radiation energy was absorbed, leading to melting enhancement.

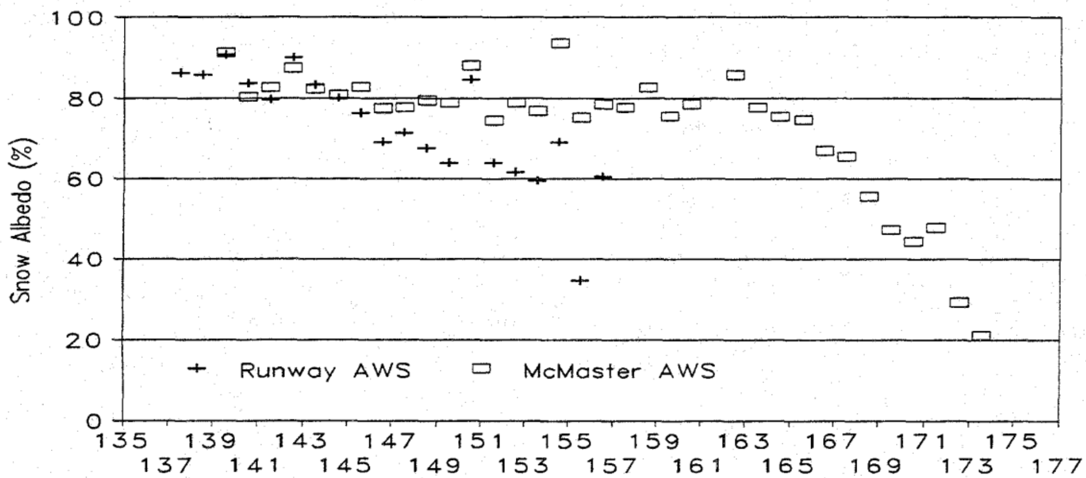

Fig. 4. Snow albedo change at dirty sand clean snow sites during the period May 15(JD 135) to June 24 (JD 175), 1994.

Snowpack Temperature

At the runway site, the snowpack was cold at the beginning of the observation period, with temperatures between -9°C and -15°C in the snowpack. The snowpack warmed up very fast from JD 141 to 147 (May 21-27) because of the sharp air temperature rise and the temperature at the middle layer of snowpack warmed by 2-3°C per day. The upper layer of the pack reached 0°C on JD 144 (May 24) and the whole pack became isothermal on JD 148 (May 28) (Fig.5).

At McMaster met. site, snow temperature was close to that at the runway site and its temperature change was similar. The upper layer warmed up to 0°C on JD 144 and the pack was isothermal on JD 150.

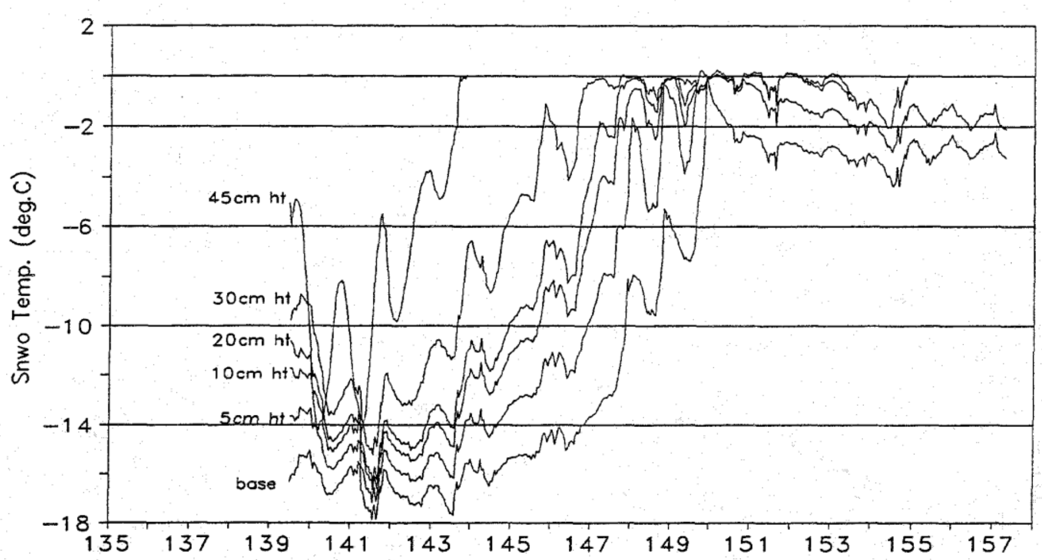

Fig. 5. Snowpack temperature at runway AWS during May 15 (JD 135) to June 6 (JD 157), 1994.

Snow Ablation

Snow ablation observation indicated that daily melt at the runway site was from 2 to 20 mm and from 2 to 14mm for the McMaster met. site. Due to the effect of dust load on snow surface albedo (Woo, 1984) and, thus, the energy received by the snowpack, there was a clear difference of snowmelt rate between the sites. At the runway site, snowpack with a higher dust load started to melt earlier and the daily melt rate was higher. As a result, the snowapck disappeared in early June. At the McMaster met. site, snowpack with a lower dust content began to melt later, the melt rate was lower, the melting period lasted longer and the snowpack persisted until late June.

The snowmelt processes at the sites was also monitored by measuring the snow depth change (Fig. 6). The decrease of snow depth was mainly due to melt at the sites in this period. The occasional increase of snow depth was due to new snowfall and blowing snow deposition. Daily snow depth measured at Resolute weather station was also plotted in Fig. 6 for comparison. It is to be noted that weather station had reported zero snow depth before the snowfall on June 1 (JD 152), while snow depth at our study sites was between 15 and 30cm. In addition, the shallow snowpack at the weather station was gone in about 12 days, compared to the melt period of 18 days versus 32 day at our study sites.

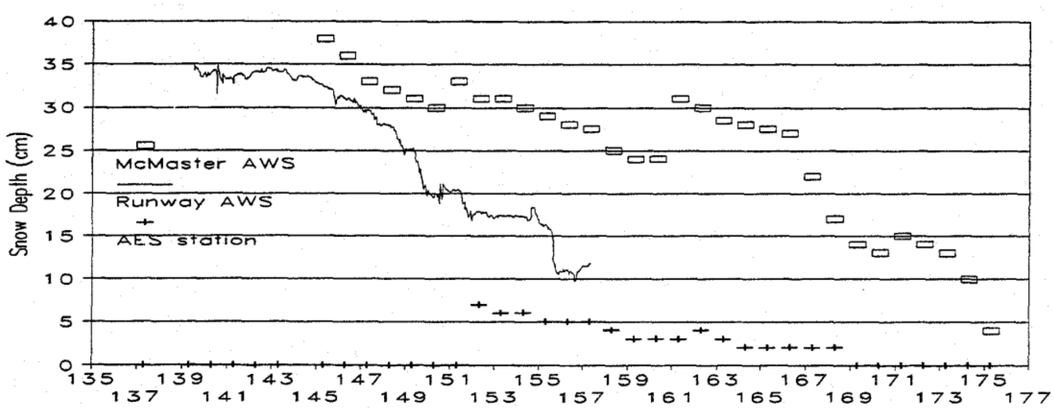

Fig. 6. Snow depth change at three sites in Resolute during May 15 (JD 135) to June 24 (JD 175), 1994.

Woo et al. (1983) found that the weather station snow data do not represent the accumulation of snow in the river basins nearby. This study suggested that caution should be taken when using the weather station snow data for regional snowcover and snowmelt studies.

CALCULATION OF SNOWMELT BY ENERGY BALANCE

The melt at the surface of a snowpack can be calculated from the energy balance

$$Q_m = Q^* + Q_h + Q_e + Q_r + Q_g \qquad (3)$$

where, Q_m is the energy available for melt, Q^* is the net radiation, Q_h is the sensible heat flux, Q_e is the latent heat flux, Q_r is the energy added by rainfall and Q_g is ground heat flux.

To determine Q_h and Q_e, the bulk transfer approach has been successfully applied to the computation of snowmelt in Arctic and subarctic regions (Price et al., 1976; Heron and Woo, 1978). In this study, the same approach was used to calculate Q_h and Q_e. Rainfall temperature was assumed to equal to the wet bulb temperature in order to estimate Q_r. Q_g was assumed to be minor important and was ignored since the ground beneath the snowpack was frozen during snowmelt period.

The roughness length of snow surface is needed to calculate Q_h and Q_e. It has been suggested that the change in roughness length (Z_0) of the snow surface was an important factor in computing snowmelt rates by energy balance method (Dunne et al., 1976; Price and

Dunne, 1976). In order to investigate the roughness length change through snowmelt period, z_0 for snow surface was calculated using the wind profile method at the runway site, where wind speed at three levels (e.g. 0.5m, 1.0m and 2.0m) was measured and hourly data were available. The analysis of the wind data indicated that hourly wind speeds at 2m were not always greater than those at 1.0m or at 0.5m. This is understandable, since the law of the increasing wind speed with height is only a statistical law, which holds good in long serial of observations, but not necessarily in individual instances. Those wind data with irregular wind profiles were identified and eliminated from the Z0 calculation. The wind data in unstable conditions were also excluded from the roughness length calculation, since air temperature gradients affect the slope of the logarithmic wind profile particularly above snow surface. The results of the calculation indicate a big variation, ranging from 2 to 30mm, of Z0 for the snow surface during the melt season. As expected, a general increase of the roughness with the progress of snowmelt was observed in late May and an unexpected sharp drop of the roughness followed in early June (Fig.7). The sharp step change of the surface roughness length for snow was obvious in Fig. 7 and may be caused by many factors, such as wind speed and the stability of the air above the snow pack. Observations with great precision of wind speed and air temperature profiles are needed to understand more the changes of the roughness of snow surface during the snowmelt period.

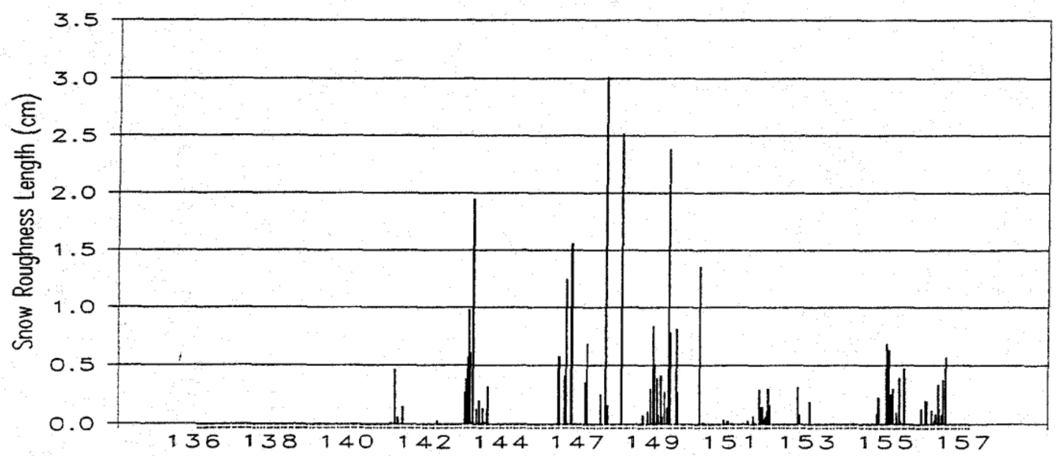

Fig. 7. Snow roughness length at runway AWS during May 16 (JD 136) to June 6 (JD 157), 1994.

The roughness length of 10mm was considered to be appropriate for the winter snow surface (Sevruk, 1982) and was widely used for reduction of wind speed from a standard height to a lower level (Sevruk, 1982; Golubev et al., 1992; Yang et al., 1993). Dunne et al. (1976) used z_0 values of 5mm and 15mm, respectively, in their energy balance snowmelt calculation on a day with high turbulent heat flow and obtained the higher melt rate for the higher z_0 input. For the energy balance calculation in this study, the medium value of 5mm of the calculated roughness length values was used.

Energy balance estimation of daily snowmelt was applied to the runway site for the period of May 16 to June 5 (JD 136 to 156) and for McMaster basin met. site for the period of May 19 to June 5 (JD 139 to 156), respectively. Loss of the snowcover and flooding of the snowmelt water at the runway site prevented calculation of melt after June 5, 1994. Fig. 8 shows the comparison of the calculated cumulative melt to the observed cumulative melt at both sites.

The energy balance estimation of the melt shows a reasonable agreement between the calculated and observed melts at both sites. The difference of melt rates at the sites was clearly reflected by the energy balance method. As expected, snowmelt began 5 days earlier at the runway site, and the melt rate was higher through the melting period. On the other hand, the calculated melt was much too high for the last 3 melt days at the runway site (Fig.8). The over-estimation of the melt might be caused by incorrect Q_h, as heat was advected from patches of ground surface near or around the met. site, and thus the sensors were not within the boundary layer above snow surface any more. In addition, high net radiation near the end of the melt season could also introduce over-estimation of the melt.

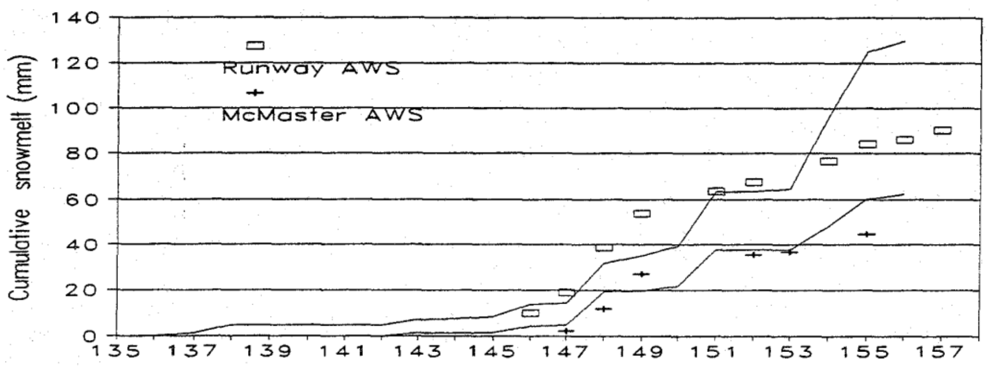

Fig. 8. Comparison of Snowmelt computed from energy balance (lines) with observed melt (symbols) at two sites during May 15 (JD 135) to June 6 (JD 157), 1994.

CONCLUSIONS

This work compared snowmelt computed from the energy balance with the observed ablation in the field. Generally, good agreement was obtained at two sites, one with high dust content and the other with relatively clean snow. This may indicate that the energy balance approach is a good short-term predictor of snowmelt in the high arctic. Uncertainties of the energy balance calculation can be attributed to the method of estimation of the snow surface roughness length.

Snow albedo decline expressed as a function of melt for different dust contents has been developed. This is useful for regional climate model and hydrological studies.

This study confirms Woo and Dubreuil's (1983) regression relation between net radiation and the product of snow albedo and the short-wave radiation. This approach can be used to estimate net radiation at weather stations, where only short-wave radiation is measured.

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