Measurements of the Energy Fluxes Involved in the Energy Budget of a Snow Cover

DONALD C. MCKAY

Atmospheric Environment Service, Downsview, Ontario, Canada M3H 5T4

GEORGE W. THURTELL

Department Land Resource Science, University of Guelph, Guelph, Ontario, Canada N1G 2W1

(Manuscript received 11 July 1977, in final form 7 November 1977)

ABSTRACT

Measurements of the various energy components involved in an energy balance of a snow cover were made at the Elora Research Station, University of Guelph during the winter of 1976. A pressure sphere anemometer and a sonic anemometer-thermometer in conjunction with fast response, fine wire, resistance thermometers and Lyman-alpha humidiometers were used to measure the fluxes of sensible and latent heat by eddy correlation techniques. A net radiometer and soil heat flux plates measured the radiative and soil heat fluxes. The data allowed a complete energy budget to be calculated including the energy stored in the snowpack and/or utilized in the fusion process, determined as a residual. The data indicated the absence of any close relationship between the net radiation and sensible and latent heat fluxes during the diurnal cycle. The flux of heat into the snowpack was found to be a major component of the energy balance. Maximum evaporation from the snow surface was observed to occur following the replacement of warm moist air masses by cold dry air masses. During these periods the sensible and latent heat fluxes were greater than the net radiation and resulted in a rapid loss of energy from the snowpack. The determination of evaporation from the snowpack using a Bowen ratio energy balance approach was found to be impractical because the large energy flux into the snowpack could not be independently determined with sufficient accuracy.

1. Introduction

Accurate information on the various heat transfer processes operative at the air-snow interface is required in such fields as agriculture, forestry, air pollution and urban studies. Snow is an important part of the hydrologic cycle in many areas of the world. Hydrologists and agriculturalists need reliable and accurate field measurements of heat transfer across the air-snow interface for their energy balance models in predicting the moisture storage of a blanket of snow and in predicting spring runoff from snowfields. In addition, with the increase of industrial activity in the arctic an urgent requirement has arisen for micrometeorological information applicable to polar regions. As well, information on the effect of regional snow cover anomalies on macroscale weather patterns is necessary before a better understanding of the general circulation of the atmosphere can be achieved. At the present time adequate field data are not available to meet the needs in these important areas.

Measurements of snow cover and its properties have been made for a century or more. Sverdrup (1936) was the first to apply the energy budget approach to a snow cover on West Spitzbergen Island in 1934. The investigation primarily dealt with the turbulent transfer of heat between the air and snow, although a complete energy balance was also computed. This work laid the foundation for most energy transfer studies over snow covers which were to follow.

The Corps of Engineers and the U. S. Weather Bureau in the 1940's initiated a cooperative snow investigation. *Snow Hydrology* (1956) was the result of 10 years of extensive data collection and analysis at three research watersheds. The results and methods included in this report are widely referenced and constitute the basis of many snow-cover models in use today. After the publication of *Snow Hydrology* and especially with the advent of the digital computer, emphasis has been placed on the development of conceptual simulation models of the snow accumulation and ablation process. Recently several models of the energy balance of a snow cover have been developed (O'Neill, 1972: Outcalt *et al.*, 1975: Anderson, 1968, 1976). The testing and development of these models require accurate experimental data, which at present are insufficient. The insufficient data have been the result of inadequate instrumentation for measuring comparatively small heat transfer values in extreme meteorological conditions. This research was designed in part to remedy the above deficiency by the acquisition of definite information on the magnitudes of the various heat budget components using the most modern instrumentation in conjunction with eddy correlation techniques. Such information, it

0021-8952/78/0339-0349$05.50
© 1978 American Meteorological Society

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

is hoped, would provide a logical basis for the further testing and development of these models.

2. Theoretical concepts

The energy balance for a snow cover can be expressed as

$$Q_N+Q_H+Q_E+Q_V+Q_G=Q_M, \tag{1}$$

where the symbols have the following meanings:

- Q_N net radiation transfer
- Q_H sensible heat transfer
- Q_E latent heat transfer
- Q_V gain of energy by all vertical advective processes (e.g., rain)
- Q_G heat transfer across the snow-soil interface
- Q_M the energy stored in the snowpack and/or utilized in the fusion process.

Eq. (1) has been applied with increasing sophistication and has been found to be generally satisfactory for synthesizing melt at a point (U. S. Corp of Engineers, 1956: Gold and Williams, 1960: Muller and Keeler, 1969: Anderson, 1968, 1976).

a. *Radiative component*

When Eq. (1) is used the radiative terms are concentrated in the single term Q_N, the net radiation. Direct measurement provides the simplest, most accurate means of obtaining this term.

b. *Turbulent transfer of sensible and latent heat*

In order to determine the significance of Q_H and Q_E to the energy balance equation it is essential that they be evaluated as accurately as possible. The terms must be calculated independently of the energy budget if the term Q_M is to be determined as a residual of the budget equation.

Q_H and Q_E may be determined directly from the eddy flux relationships

$$\left.\begin{aligned} Q_H&=-C_p\overline{\rho w'T'}\approx -C_p\rho\overline{w'T'}\\ Q_E&=-L_s\overline{\rho w'q'}\approx -L_s\rho\overline{w'q'}\end{aligned}\right\}, \tag{2}$$

where the symbols are as follows:

- C_p the specific heat of air at constant pressure
- L_s the latent heat of vaporization/sublimation
- T' the fluctuating component of temperature, T, defined by the relationship $T'=T-\bar{T}$
- q' the fluctuating component of specific humidity, q, defined by the relationship $q'=q-\bar{q}$
- w' the fluctuating component of the wind normal to the earth's surface; $w'=w$ when $\bar{w}=0$, is assumed
- ρ the air density

and the overbar indicates a time average.

Instruments for eddy correlation measurements based on Eqs. (2) and (3) have been developed to measure sensible heat and water vapor flux (Dyer, 1961: Tanner and Thurtell, 1969). It is not practical to measure these fluxes by eddy correlation instruments for routine hydrologic applications (Anderson, 1976). However, the eddy correlation method was adopted for this study because of the availability of these instruments and the required expertise for their operation, coupled with the desire to obtain the most accurate values of Q_H and Q_E.

c. *Advective energy contribution*

The advective term Q_v, or the heat transfer by mass changes includes the energy received from rain, the vapor transfer between the snow and air and snow and soil and the latent heat resulting from freezing of rainwater on contact with subfreezing snow. O'Neill (1972) found that mass flux resulting from evaporation or condensation is small and the precipitation term is the only term to be considered. As no data were collected either during or for some time after a period of rain, Q_v was taken to be zero in this study.

d. *Soil heat transfer*

Accurate evaluation of the heat flux across the soil-snow interface is complicated by the effects of many properties and processes, for example, soil moisture content, infiltration of melt water, vapor transfer and the magnitude of solar radiation penetration through the overlying snowpack. Several approaches have been attempted to measure Q_G in soils. Heat flow plates have been buried in the soil (Monteith, 1958; Selirio, 1975) or soil temperature profile measurements have been made in conjunction with estimates of soil heat capacity to determine net gain or loss of heat within a soil layer near the surface (Slayter and McIlroy, 1961). In the studies reported here the direct approach, that is, heat flow plates buried just under the soil surface were used to measure Q_G directly.

e. *Storage and/or fusion component*

When all the energy components appearing on the left hand side of Eq. (1) are evaluated the value remaining Q_M constitutes the energy stored in the snow cover. This energy may remain in the snowpack as stored heat or it may be utilized in the fusion process. Q_M is the most difficult of all the energy component terms to to measure directly. In order to measure Q_M a number of snowpack parameters, such as the density and temperature of the solid portion of the snowpack and the liquid water content, must be known precisely. In regions where thawing and refreezing occur a number of times during the season the snow cover parameters are constantly changing. This makes direct measurements of the properties of the snowpack nearly impossible. For this reason Q_M is usually determined as

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

the residual term in the energy budget and this procedure was adopted for this study.

3. Description of the site

The experiments were conducted in a nearly flat 25 ha field at the Elora Research Station (43°39′N and 80°35′W), about 23 km north of the University of Guelph, in the winters of 1975 and 1976. The field is planted in corn (*Zea mays*) during the summer and is fall ploughed after harvesting. It is surrounded by further large experimental fields, slopes no greater than 1 to 2%. The nearest obstructions from the experimental site are a barn and silo towers about 500 m to the north and a service storage building about 700 m to the east. The instrument trailers are located on the east side of the measurement area and a climatological station is located to the northeast (Fig. 1). The site is quite adequate for the micrometeorological studies reported.

In 1976, three instrument towers were installed as shown in Fig. 1. Tower 1 and tower 2 supported the pressure sphere and sonic anemometers, respectively, while tower 3 supported the net radiometer. The towers supporting the anemometers had a wind fetch of over 400 m except when the wind was from the east at which time it was reduced to 70 m as the wind would be over the instrument trailers. During these periods no data were collected.

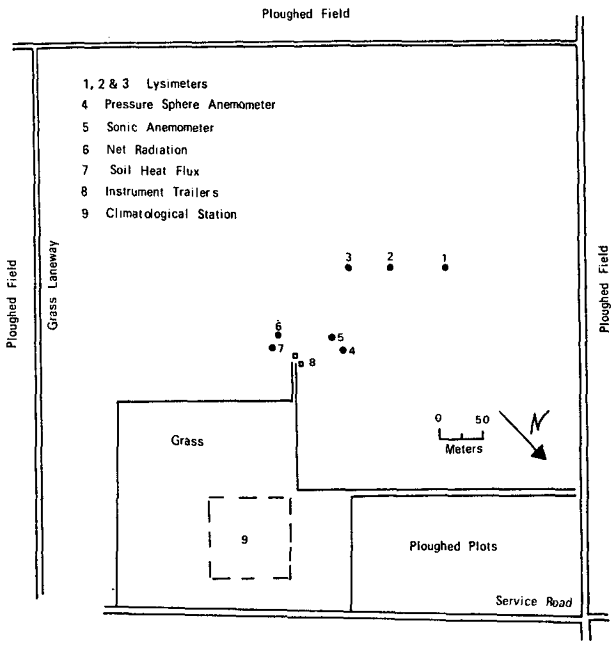

FIG. 1. Agrometeorology research area at the Elora Research Station, 1976.

4. Experimental methods

a. Wind velocity measurements

Two independent anemometers were used to measure the three components of wind velocity. The first was a pressure sphere anemometer described by Thurtell *et al.* (1970) and Wesely *et al.* (1972), the second was a sonic anemometer thermometer (Kaijo Denki)[1] described by Mitsuta (1966). The instruments were mounted on towers 21 m apart, the pressure sphere anemometer at a height of 3.7 m and the sonic anemometer at 3.5 m above the surface. As measurements in the atmosphere are difficult to verify because no accepted standard technique is available, confidence in the various methods can be achieved by measuring the same quantity independently. For this reason the two independent anemometers were used. Wind direction was obtained from a wind vane at the 10 m height situated at the climatological station. Also a vane was attached to the pressure sphere anemometer which allowed it to align itself with the mean wind.

b. Temperature measurements

Temperature fluctuations were measured with a fast response, fine-wire resistance thermometer (Wesely *et al.*, 1969; Silversides, 1972). The thermometer consists of an approximately 0.4 m length of 5.6 μm diameter platinum-coated tungsten wire wound on an insulated metal frame. The change in resistance of the thermometer is detected by means of a bridge, the design and electronics of which are described in detail by Silversides (1972).

Two of these resistance thermometers were used, one in conjunction with the pressure sphere anemometer, the other in conjunction with the sonic anemometer. As well as the resistance thermometers, the temperature sensor inherent in the sonic anemometer-thermometer also was monitored. This allowed two estimates of Q_H to be made from Eq. (2) by combining the vertical wind fluctuations detected by the sonic anemometer first with data from a resistance thermometer, and second with data from the sonic thermometer.

The surface temperature of the snow cover was monitored by an infrared thermometer (Barnes Engineering Model 1T-4)[2] which was focused on the snow site through an opening in one of the instrument trailer windows. Air temperature was detected using a Rosemount resistance temperature sensor (Rosemount Engineering Co.)[3] located in a Stevenson screen at the climatological station.

c. Water vapor measurements

Two Lyman-Alpha Humidiometers (Electromagnetic Research Corporation),[4] hereafter referred to as Lα were used to measure the humidity fluctuations over the

[1] Kaijo Denki Co., Ltd., Tokyo, Japan.

[2] Barnes Engineering, 30 Commerce Rd., Stamford, Conn., U. S. A.

[3] Rosemount Engineering Co., Minneapolis, Minn., U. S. A.

[4] Electromagnetic Research Co., College Park, Md.

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

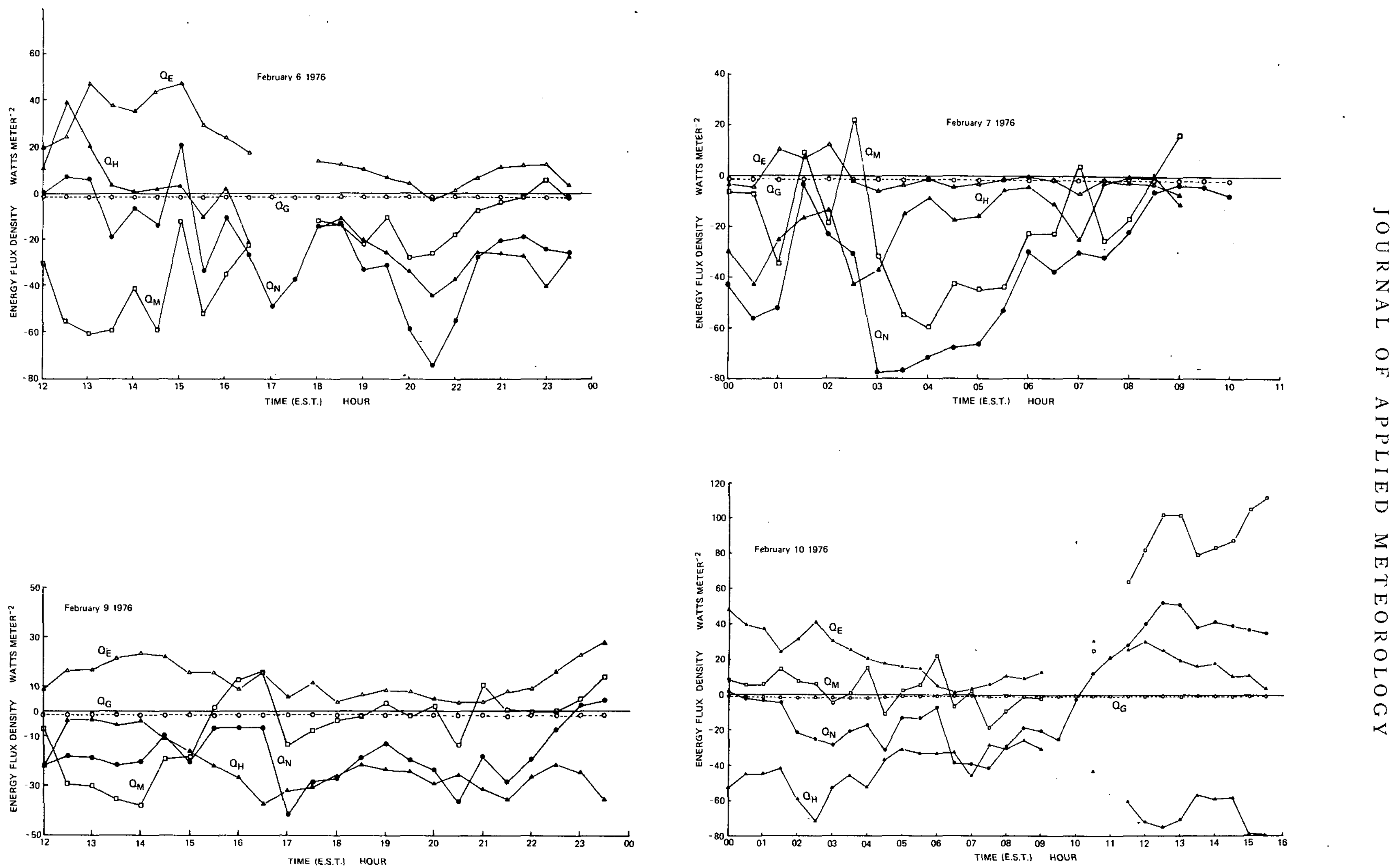

FIG. 2. Diurnal trends of net radiation, Q_N, (●—●—●); sensible heat flux density, Q_H, (▲—▲—▲); latent heat flux density, Q_E, (△—△—△); soil heat flux density, Q_G, (○—○—○); and energy stored in the snowpack and/or utilized in fusion process, Q_M, (□—□—□), for 6 February 1976 (a), 7 February 1976 (b), 9 February 1976 (c), 10 February 1976 (d), 12 February 1976 (e), 13 February 1976 (0000–1500) (f), 13 and 14 February 1976 (g), 14 February 1976 (0600–1800) (h), 25 February 1976 (i), 17 March 1976 (j), 19 March 1976 (k) and 20 March 1976 (l).

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

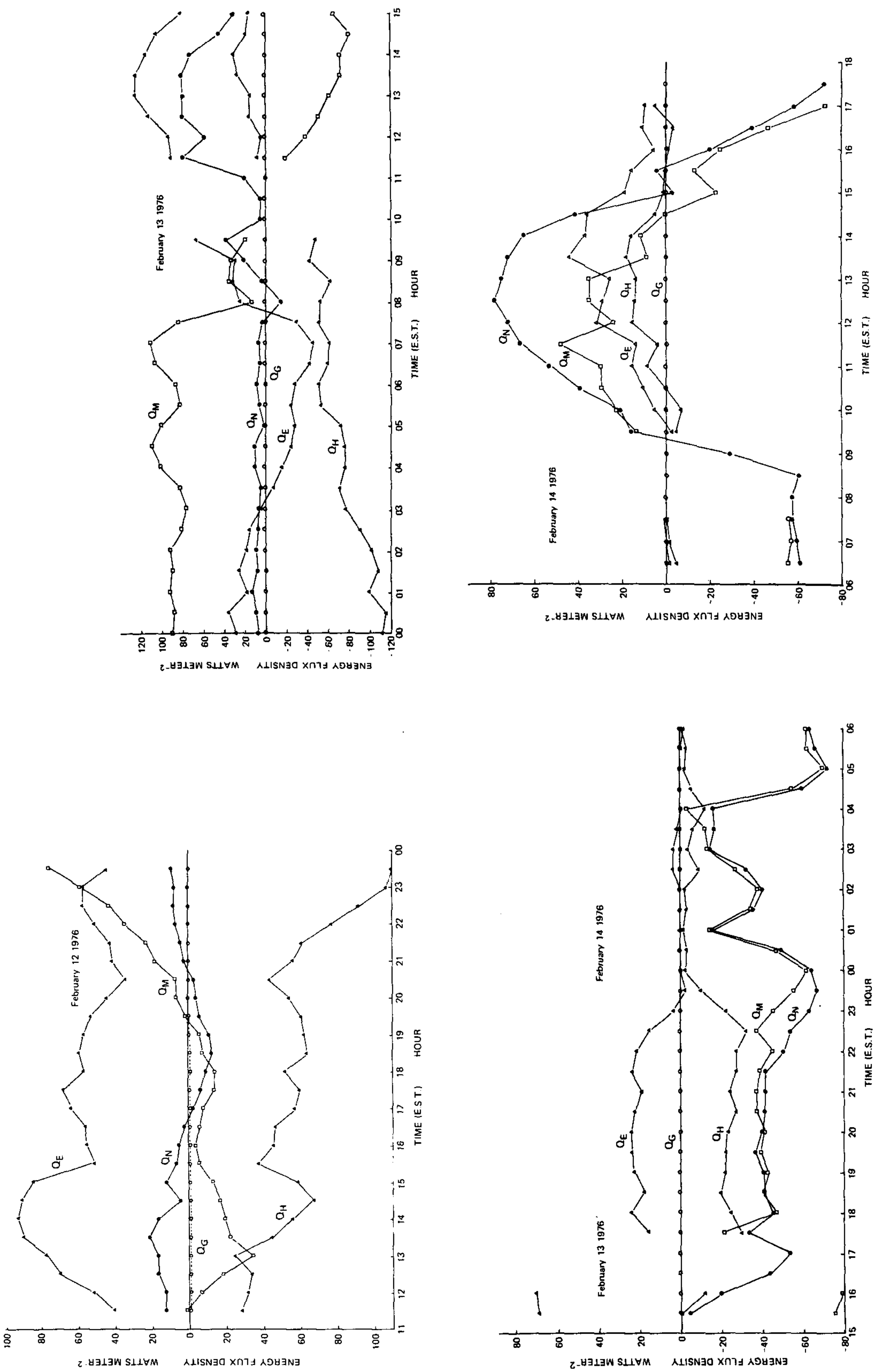

FIG. 2. (*Continued*)

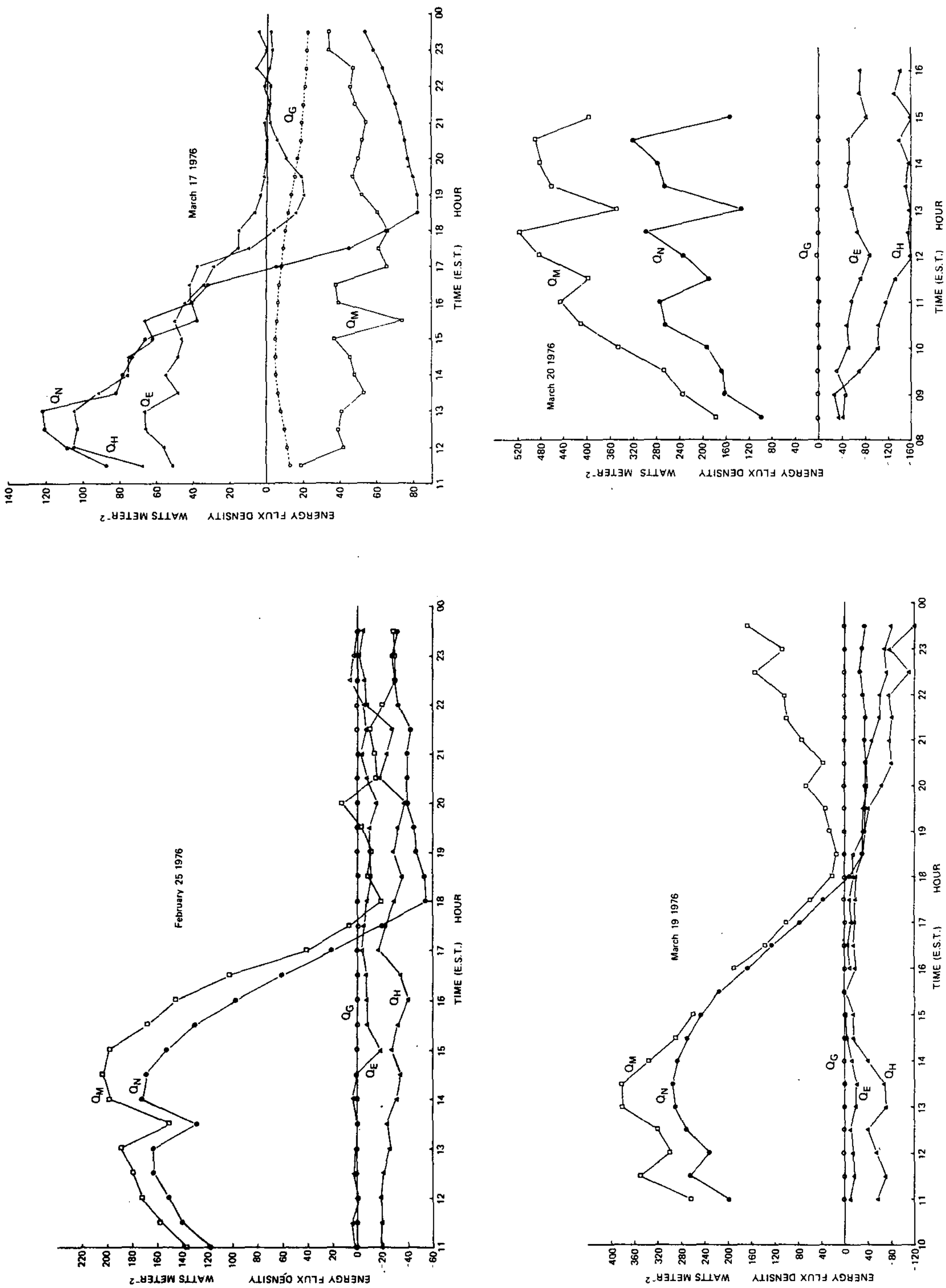

FIG. 2. (*Continued*)

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

TABLE 1. Synoptic weather conditions.

6–7 February	Cold continental air mass over the region, winds westerly.
9 February	High-pressure ridge east of region with warm front approaching from the west, southwesterly winds.
10 February	Warm air being advected into the area ahead of TROWAL extending across Lake Huron and Georgian Bay.
12 February	Warm front running west to east across the lower Great Lakes just south of site.
13 February	Cold frontal passage through the region at approximately 0800 EST. Cold continental Arctic air moving into the region.
14 February	High-pressure area covering the region, sunny skies with cold temperatures.
25 February	Warm front approaching from south, southerly winds advecting warm air into the region.
17 March	Cold continental air mass flowing into the region after cold frontal passage on 16 March.
19 March	Quasi-stationary front lying just north of the area, warm air being advected in by strong southerly winds.
20 March	Warm front lying across Georgian Bay to north of region. Warm maritime polar air mass covering the area.

snow cover. One was attached to the sonic anemometer the other to the pressure sphere support. As the humidiometers were used to measure humidity fluctuations no mean humidity values were calculated from these instruments. The Lα's were operated in conjunction with the sonic and pressure sphere anemometers and temperature sensors in order to obtain simultaneous values of Q_E and Q_H at the two sites.

The dewpoint of the air was continuously monitored using a Lithium-chloride dewcell (A.E.S. type) located in a Stevenson screen at the climatological stations.

d. *Net radiation measurements*

Net radiation over the site was obtained using an SRI[5] net pyrradiometer installed at a height of 3 m above the surface. The pyrradiometers are commercially available and are capable of accuracies of 5% or better on a daily time scale, when properly operated and maintained (Latimer, 1972).

e. *Soil heat flux measurements*

Soil heat flux measurements were obtained over the site using four soil heat flow plates (Thornthwaite Assoc., Elmer, N.J.) connected in series and buried approximately 1–2 mm under the soil. The heat flow plates were calibrated in the laboratory by generating a known one-dimensional heat flow as described in detail by King (1974).

f. *Data collection system*

The wind signals from the pressure sphere anemometer along with the temperature signals from the resistance thermometer bridge amplifier were recorded directly by the computer. The sonic anemometer-thermometer signals and the signals from the Lα's were conditioned by Neff type 119 amplifiers with a bandwidth of 100 Hz. The net radiation, soil heat flux and infrared thermometer signals were conditioned by Prestion X-Mod 706 amplifiers with a bandwidth of 10 Hz. Each signal was amplified to ± 10 V before being recorded by the computer.

Signals were analysed "on line" by a Honeywell 316 digital computer (Honeywell Controls Ltd.)[6] housed in the mobile trailer. Outputs from all anemometers, thermometers, Lα's, radiometers and soil heat flow plates were sampled one hundred times per second. For the eddy correlation calculations each of five parameters (u,v,w,T,q) was calculated at each sampling point together with its square and all cross products. At the end of each 30 min sampling period, all means, variances and covariances were computed. A coordinate transformation to make $\bar{v}=\bar{w}=0$ was applied as described by Tanner and Thurtell (1969) and the results including the sensible heat, latent heat and momentum fluxes were printed out and punched onto paper tape.

5. Results and discussion

Sample results of direct measurements of the energy balance components for various weather conditions (Table 1) are shown in Figs. 2a–2l. The values of Q_H and Q_E are average values. Three sets of data were averaged to obtain Q_H (one value for each thermometer) and two sets of data were averaged to obtain Q_E (one value for each Lα). The value of Q_M was obtained by using the above average values of Q_H and Q_E along with Q_N and Q_G. The illustrated examples consist only of observation periods on which fairly long records of continuous measurements were obtained. 6 and 7 February and 17 March represent time periods when a cold continental air mass was covering the area. 9 and 10 February represent periods when a warm air mass was advected in to replace a cold air mass. 12, 13 and 14 February represent a transition period during which a cold frontal passage occured (13 February) bringing in a cold air mass to replace a warm air mass that was over the area. 25 February and 19 and 20 March are examples of periods when extensive snow melt was occurring. Three features of the data are particularly striking:

1) The absence of any diurnal periodicity in the sensible and latent heat flux terms.

2) The control of either the radiation or the sensible heat flux upon Q_M, depending upon the meteorological condition.

3) The occurrence of maximum evaporation during the transition period from a warm to a cold air mass.

[5] Solar Radiation Instruments, Victoria, Australia.

[6] Honeywell Controls Ltd., 740 Ellesmere Rd., Scarborough, Ontario.

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

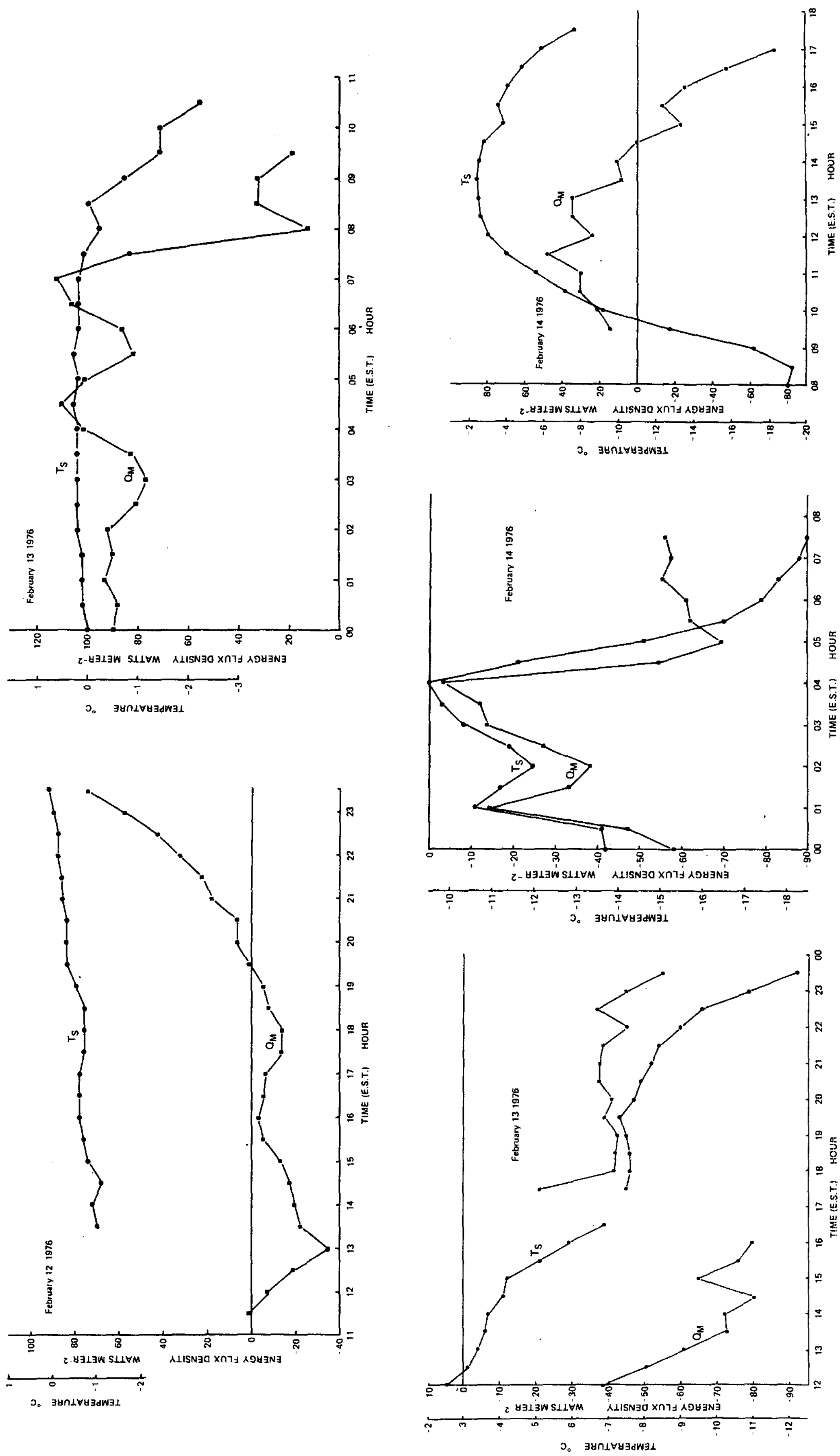

FIG. 3. Diurnal trends of the energy flux density of Q_M, (■—■—■), and the surface temperature T_s, (●—●—●), for 12 February 1976 (a), 13 February 1976 (0000–1100) (b), 13 February 1976 (1200–0000) (c), 14 February 1976 (0000–0800) (d) and 14 February 1976 (0800–1730) (e).

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

In energy budget studies conducted in summertime conditions over bare soil, over grass or over a field crop there is a close relationship between the sensible heat and latent heat fluxes and the net radiation (Yap and Oke, 1974; Tanner and Thurtell, 1969). Maxima and minima in the diurnal course of Q_N generally result in a corresponding behavior in Q_H and Q_E. This relationship was not often found over the snow cover. During the winter a large portion of the energy stored in the snowpack does not come from the net radiation but from warm air masses which have been advected into the region by macroscale weather systems. Thus, when a cold dry air mass moves back in over the region this stored energy is released as sensible and latent heat which is not related to the net radiation. This is exhibited in all the figures but particularly on 12 and 13 February, 25 February and 19 and 20 March. The turbulent terms do appear to be important to the energy exchange during particular periods, for example, by partially compensating for radiative heat loss overnight (7, 10, 13 February and 17 March). This is significant in that the initiation of melt following a period of net energy loss is dependent upon the "cold content" of the snowpack (O'Neill, 1972), the "cold content" being the temperature that is attained by the snowpack during a period of net energy loss. As the "cold content" decreases less heat is required to raise the temperature of the pack to 0°C and a proportionately increased percentage of the energy supply becomes available for the fusion process.

The control of the net radiation or the sensible heat flux on the fusion process appears to depend upon the meteorological conditions. In cases where the fusion process is controlled by the net radiation, (9 February, 19 March 1100–1800 E.S.T., 20 March) the air mass over the region is well established and there is little daily variation in the air temperature. When the air over the region is in the process of change, for example, when warm air is being advected in to displace the cold air (12 February and 14 February) the sensible heat flux term, Q_H, then becomes the controlling mechanism for the fusion process. This would suggest that temperature index models, which are formulated on the concept that sensible heat exchange is the primary source of energy for melt, cannot be used without discretion for estimating snowmelt. The data suggest that the energy balance approach would be more reliable in that variability in meteorological conditions are reflected in the different energy components which are not necessarily detected by the temperature index approach. This has been noted by Anderson (1976) in his testing of the two types of approaches.

The data presented lend credence to the fact that maximum evaporation occurs during periods when a cold air mass is displacing a warm air mass which has caused snowmelt. For example, on 12 and 13 February during the periods 2000–0300 E.S.T., the average Q_E was 35.3 W m^{-2}. Snowmelt was occurring at this time as indicated by the large positive Q_M values (Figs. 2e and 2f). Also a warm air mass was situated over the area as noted in Table 1. On 13 February at approximately 0800–1600 EST the average Q_E value was 81.7 W m^{-2}. A possible explanation for this observation is that when a warm air mass intrudes the region it is generally moist. Therefore, most of the available energy transported to the snowpack is used to warm the snowpack up to 0°C. When this plateau is attained, the remaining energy goes into the melt process. The cold air masses moving into the region are relatively dry while the snowpack is moist with melt water. Thus there is a large vapor pressure gradient developed at this time between the snow surface and the air and increased evaporation occurs.

It is interesting to note the relationship between the surface temperature T_S and the energy component Q_M. Fig. 3 shows plots of Q_M and T_S against time for the period 12 to 14 February. When Q_M is negative implying the snowpack is losing energy one would expect the surface temperature to decrease. When Q_M is positive implying the snowpack is gaining energy, one would expect the surface temperature to increase. This hypothesis is verified by the data shown as indicated by time periods 0800–0000 on 13 February (Figs. 3b and 3c) and 0000–1800 on 14 February (Figs. 3d–3e). However, there is an upper limit to which the snow surface temperature can attain, that being 0°C. After the snow surface or the snowpack has achieved this upper limit any further energy gains must go into melting the snowpack. One exception to this relationship is noted during the period 1200–1900 EST on 12 February. During this interval Q_M is negative indicating that the surface temperature should be decreasing. The data indicates that the surface temperature is increasing (Fig. 3a). A possible explanation for this occurrence is vapor diffusion up through the snowpack. If one examines the weather conditions two days prior to this period it will be noted that on 10 February melt was occurring followed by a refreezing interval on 11 February. On 12 February during the period in question an ice crust had formed on the snow surface. If the snow beneath this surface was moist due to snowmelt infiltration on 11 February, then water vapor could move upward through the pack from the warmer levels below. This water vapor diffusion from below could account for the relatively higher values of Q_E recorded during this time (Fig. 2e). Also some of this vapor could have condensed out on the cooler upper surface releasing latent heat and thus warming the ice crust.

Measuring the turbulent terms Q_H and Q_E by eddy correlation techniques is technically difficult and expensive. From a practical point of view it is more useful if one can relate these fluxes to more standard observations such as mean wind speed, snow-air tem-

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

TABLE 2. Daily average Bowen ratios calculated by gradient and eddy correlation methods.

Date	Bowen ratio Gradient	Eddy correlation
9/2/76	−4.1	−2.5
12/2/76	−1.1	−1.0
13/2/76	−1.9	−0.5
25/2/76	2.5	3.9
19/3/76	1.7	2.8
20/3/67	1.2	2.0

perature and humidity differences. The energy balance Bowen ratio method has been widely used in micrometeorological research because of its simplicity in theory and in instrumentation requirements. In order to investigate whether a more operational method could be used to obtain the turbulent fluxes over the snow cover a comparison of Bowen ratios was undertaken. Since the eddy correlation method measured Q_H and Q_E independently a Bowen ratio using

$$B_e=\frac{Q_H}{Q_E} \tag{4}$$

was calculated. A second Bowen ratio using the gradient method was obtained using the formula

$$B_g=\gamma\frac{T_a-T_s}{e_a-e_s}, \tag{5}$$

where the symbols are as follows:

- γ $C_P P/L_s\epsilon$
- C_P specific heat of air at constant pressure
- P average air pressure
- L_s latent heat of sublimation
- ϵ ratio of molecular weights of water vapor to air
- T_a temperature of the air
- T_s temperature of the snow surface
- e_a vapor pressure of the air
- e_s vapor pressure of the snow surface which was taken to be the saturated vapor pressure at T_s.

The procedure used in calculating the Bowen ratio by the gradient method was to first obtain a ½ h value from ½ h averaged temperature and vapor pressure measurements and then average the ½ h Bowen ratio values over the day.

Table 2 shows the daily averages for the Bowen ratio calculated by the two methods. As the data indicate the values differ somewhat between the two methods except for 12 February where there is remarkable agreement. The difference in values between the two methods could be caused by a number of factors.

In calculating the Bowen ratio by the gradient method one must know the dewpoint of the air in order to determine e_a. A lithium-chloride dew cell was used to obtain the dew point. At temperatures below 0°C the value of the dew point can be in error by as much as 2°C when measured by this instrument. This could introduce a percentage error of 16% in the value of the vapor pressure of the air. Also errors could be introduced in the measurement of e_s, the vapor pressure of the snow surface. To obtain this value the assumption made was that the surface temperature and the dew-point temperature of the snow surface are equal. The surface temperature was monitored by an infrared thermometer. As Fuchs and Tanner (1966) note, errors up to 1°C can be introduced depending upon how well the emissivity of the surface is known. Thus an error in the surface temperature would also introduce an error into the gradient method calculation of the Bowen ratio. However, this error is expected to be minimal due to the fact that in the operational wavelength band of the infrared thermometer (8–12 μm) the emissivity for snow is very close to 1.0. This is indicated in Figs. 3a and 3b where the surface temperature is hovering near 0°C while snowmelt is occurring and the air temperature during this time was well above freezing.

Another aspect worth considering in this section is the feasibility of using Bowen ratio calculations for energy budget determinations in winter time situations. As seen in the data presented in Figs. 2a to 2l the sensible and latent heat terms are relatively small compared to summer time values. In most cases one finds that in the winter time the fluxes of sensible and latent heat are about an order of magnitude smaller than their summer counterparts. Because of these small values the errors introduced in measuring them will increase and thus the error in the Bowen ratio will be increased. Furthermore, during the summer the energy available for sensible and latent heat comes predominantly from the net radiation, in winter much of the latent heat is transported from the air mass as sensible heat. The Bowen ratio energy balance approach assumes that the energy available comes from measurable energy sources for example, the net radiation, energy stored in the surface. In winter the energy stored in the snowpack is usually the source of sensible and latent heat. To use the Bowen ratio approach Q_M must be known. However, in most practical situations Q_M is determined as the residual term from the energy balance due to the difficulty experienced in trying to measure it. Therefore, for these reasons the Bowen ratio energy balance approach must be considered inadequate for precise energy balance measurements during winter time conditions.

6. Summary

The research has examined the energy balance for a snow cover. The data indicate that neither latent nor sensible heat fluxes display a marked diurnal periodicity as found in the summer. This relationship is not found over the snow cover because much of the latent heat energy is transported from the airmass as sensible heat

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC

and is not obtained from the net radiation. Also, the components controlling the energy storage and/or fusion process depended upon the meteorological situation. The net radiation was the controlling component during periods when the air mass was well established over the region, while the sensible heat component became the controlling term when warm air was advected into the region. These facts suggest that an energy balance approach to predicting snowmelt is favored over a temperature index model since the former will reflect the variability in the meteorological conditions while the latter may not detect this variability. The maximum evaporation from the snow surface occurs during transition periods when a cold air mass flows into the region replacing a warmer air mass which has produced melting conditions.

The Bowen ratio energy budget approach appears to be less applicable to wintertime conditions as compared to summertime conditions. This is due to the relatively small values of the turbulent components of latent and sensible heat in winter as compared to summer (usually an order-of-magnitude difference). Also the net radiation is not necessarily the source of energy for sensible and latent heat during the winter. In most situations in winter the source of energy is the energy which has been stored in the snowpack during periods when a warm air mass has been advected into the region. Therefore, the Bowen ratio energy balance approach breaks down during wintertime conditions since the large energy flux into the snowpack cannot be independently determined with sufficient accuracy.

Acknowledgments. The first author is indebted to the Atmospheric Environment Service for granting him educational leave and providing financial support to undertake this study. Both authors wish to thank Mr. G. Kidd for his assistance with the computer programming throughout the project.

REFERENCES

Anderson, E. A., 1968: Development and testing of snow pack energy balance equations. *Water Resour. Res.*, **4**, 19–37.

——, 1976: A point energy and mass balance model of a snow cover. NOAA Tech. Rep. NWS 19, 150 pp. [Available from Superintendent of Documents, U. S. Govt. Printing Office, Washington, D. C.]

Dyer, A. J., 1961: Measurements of evaporation and heat transfer in the lower atmosphere by an automatic eddy correlation technique. *Quart. J. Roy. Meteor. Soc.*, **87**, 401–412.

Fuchs, M., and C. B. Tanner, 1966: Infrared thermometry of vegetation. *Agrometeor. J.*, **58**, 597–601.

Gold, L. W., and G. P. Williams, 1960: Energy balance during the snow melt period at an Ottawa site. *IASH Gen. Assembly of Helsinki*, 288–294 pp.

King, K. M., 1974: Meteorological instrument calibration. Guelph Project Final Rep. 1974, Training in agrometeorology and research on photosynthesis of crops in relation to productivity. Canadian Committee I.B.P. 93–95. [Available from Department of Land Resource Science, University of Guelph, Guelph, Ontario].

Latimer, J. R., 1972: Radiation measurement. Int. Field Year on the Great Lakes, Tech. Manual Series, No. 2, Secretariat, Can. Nat'l Committee for the IHD, No. 8 Bldg. Carling Avenue, Ottawa, 53 pp.

Mitsuta, Y., 1966: Sonic anemometer-thermometer for general use. *J. Meteor. Soc. Japan*, **44**, 12–23.

Monteith, J. L., 1958: The heat balance of soil beneath crops. *Proc. Symp. UNESCO Arid Zone Res. Climat. Microclimat.*, Canberra, No. 11, 123–128.

Muller, F., and C. M. Keeler, 1969: Errors in short term ablation measurements on melting ice surfaces. *J. Glaciol.*, **8**, 99–105.

O'Neill, A. D. J., 1972: The energetics of shallow prairie snow packs. Ph.D. thesis, University of Saskatchewan, 197 pp. (Available from National Library, Ottawa, Canada).

Outcalt, S. I., C. Goodwin, G. Weller and J. Brown, 1975: A digital computer simulation of the annual snow and soil thermal regimes at Barrow, Alaska. Cold Regions Research and Engineering Laboratory Research Rep. 331, Hanover, N. H. 18 pp.

Selirio, I. S., 1975: Study of variation of apparent Bowen ratio with height above a cornfield. Ph.D. thesis, University of Guelph, 77 pp. [Available from National Library, Ottawa, Canada].

Silversides, R. H., 1972: The structure of thermal turbulence above and within canopies of *Zea mays* (L.) and Pinus resinova (AIT). Ph.D. thesis, University of Guelph, 262 pp. [Available from National Library, Ottawa, Canada.]

Slayter, R. O., and I. C. McIlroy, 1961: *Practical Microclimatology.* CSIRO, Australia, 250 pp.

Sverdrup, H. U., 1936: The eddy conductivity of the air over a smooth snow field—results of the Norwegian Swedish Spitzbergen Expedition in 1934. *Geofys. Publik.*, **11**, 1–69.

Tanner, C. B., and G. W. Thurtell, 1969: Anemoclinometer measurements of Reynolds stress and heat transport in the atmospheric surface layer. ECOM 66-F 82 pp. [Available from U. S. Army Electronics Command, Fort Huachuca, Ariz.]

Thurtell, G. W., C. B. Tanner and M. L. Wesely, 1970: Three-dimensional pressure-sphere anemometer system. *J. Appl. Meteor.*, **9**, 379–385.

U. S. Army Corps of Engineers, 1956: *Snow Hydrology.* Summary Report of the Snow Investigations. North Pacific Div., Corps of Engineers, Portland, Ore., 437 pp.

Wesely, M. L., G. W. Thurtell and C. B. Tanner, 1969: A fast response thermometer for eddy correlation measurements. In Anemoclinometer measurements of Reynolds stress and heat transport in the atmospheric surface layer, ECOM 66-G22-F, pp. 61–72.

——, C. B. Tanner and G. W. Thurtell, 1972: An improved pressure sphere anemometer. *Bound.-Layer Meteor.* **2**, 275–283.

Yap, D., and T. R. Oke, 1974: Eddy-correlation measurements of sensible heat fluxes over a grass surface. *Bound.-Layer Meteor.*, **7**, 151–163.

Unauthenticated | Downloaded 05/31/23 05:14 PM UTC