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ENERGY BALANCE OVER MELTING SNOW, CRAIGIEBURN RANGE, NEW ZEALAND

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ABSTRACT

Over a total of thirty-one studied days of spring melt between 1975 and 1980, sensible heat was the major source of energy to the snowpack, contributing approximately sixty percent of the total heat supply. Net radiation was second most important, although on some days of heavy cloud, high humidity and windiness, latent heat flow exceeded it. Precipitation heat flow and ground heat flow were both relatively unimportant. However, the greatest heat flows occurred on days with rain, mainly because such days were also warm and windy with high sensible heat transfers.

INTRODUCTION

Modelling of runoff from snow requires information on (a) energy available for melt, (b) areal extent of snow and (c) routing of snowmelt from snow surface to channels. Although the Technical Subcommittee on Snow of the International Hydrological Decade (1969) recommended that investigations should be undertaken, a decade later, Fitzharris (1979) observed that little progress in these areas had been made in New Zealand.

Of the three topics, no New Zealand studies of the routing of snowmelt are known to the authors, and the data relating to the extent of snow are rather sparse. The ratio of snow courses in New Zealand to the total area of winter snow is about 1 to 4,000 km^2, while within the San Juan Mountains of the United States the ratio is only 1 to 268 km^2 (Caine, 1975) and it is even lower at 1 to 77 km^2 within the Snowy Mountains of Australia (Snowy Mountains Hydro-Electric Authority, 1970).

Several studies of energy available for melting have been made on glacier surfaces in New Zealand (Anderton, 1976a, b; Anderton and Chinn, 1978; Dickson, 1974; and Harding, 1972) but Fitzharris *et al.* (1980) appear to be the only ones to have calculated heat flows for a melting seasonal snowpack. Even this work, however, required extrapolations of meteorological parameters and was limited to a single three-day event. Energy balance measurments for a high location in Canterbury have also been reported by Greenland (1973), though snow had negligible influence at the site studied.

Many energy balance studies have been made of melting snowpacks overseas, but often in conjunction with the development of operational runoff models (Anderson, 1973). As a result, presentation of heat-flow data is limited. However, some generalizations can be drawn from the available results. Net radiation as a dominant heat source has been frequently reported, especially for forested areas (U.S.A.C.E., 1956, 1960; Hendrie and Price, 1979) and at high altitudes in mountainous areas (de

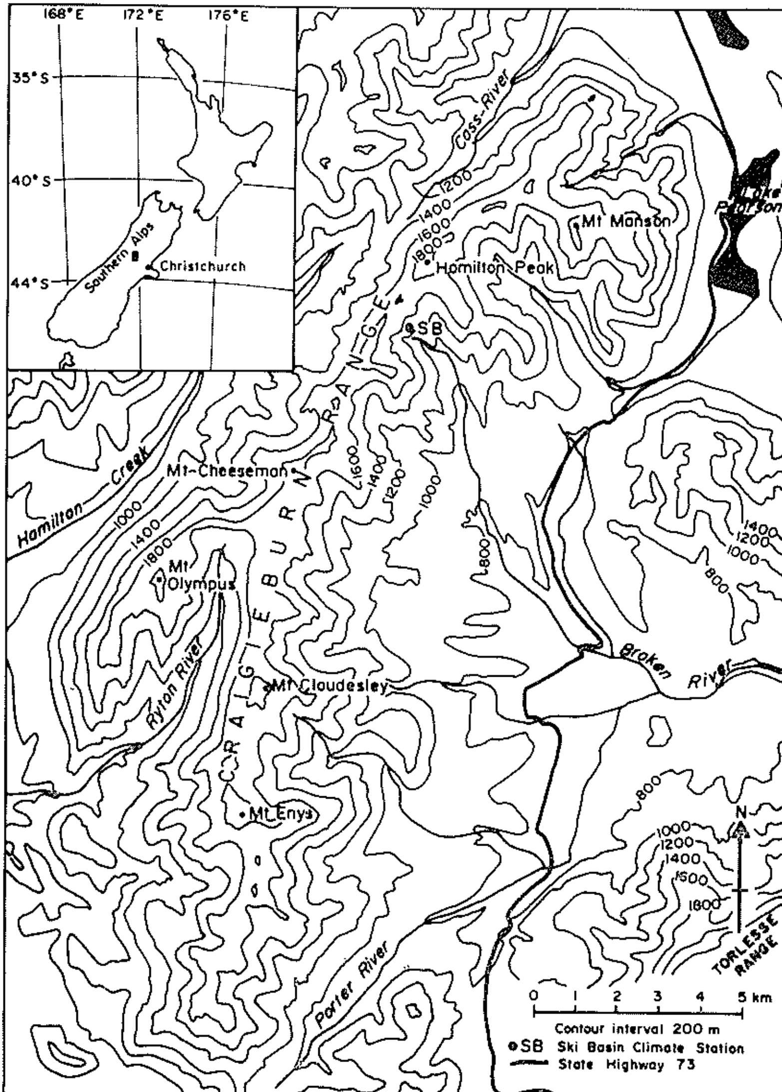

FIG. 1—Location Map.

La Casiniere, 1974; Jordan, 1978). In other studies, for example those at lower-altitude treeless locations in Canada (Granger and Male, 1978; Price, 1975; McKay and Thurtell, 1978) turbulent transfers of energy play a much more important role which varies with the passage of different meteorological situations.

This paper presents results of the application of the energy balance method to periods of spring snowmelt in the Craigieburn Range, New Zealand (Fig. 1). It assesses the relative importance of the individual heat flow components during both long-term and intense periods of seasonal snowmelt.

Site Description

Data for this study was collected in the Broken River catchment of the Craigieburn Range, 20 km east of the main divide of the Southern Alps (Fig. 1). A number of peaks in the range exceed 2000 m but the average ridge elevation is approximately 1800 m. The topography is characterized by rounded ridge crests and slopes of 30 to 40° above 1200 m. The lower slopes are covered in part by mountain beech forest and the higher elevations are dominated by snow tussock and scree.

The Craigieburn Range lies perpendicular to the path of the prevailing westerlies approximately 30 km downwind of the point of maximum precipitation (McSaveney, 1978). Of the 1800 mm of annual precipitation received at 1500 m, over 80 percent originates from the westerly quarter (McCracken, 1980). Snow may fall at any time of the year and contributes approximately one third of the annual precipitation above 1500 m but only negligible quantities below 900 m. Most snow accumulates above

FIG. 2—Ski Basin climate station. The Alter shield was fitted to the precipitation gauge in 1980.

treeline (1380 m), usually beginning in May or June and reaching a maximum in September or October (O'Loughlin, 1969).

At 1500 m, July is the coldest month, with a long-term mean air temperature of $^-1.4$°C, and February the warmest at 9.7°C. The mean monthly temperature usually rises above freezing in August or September. The annual temperature range is not as great as found in continental climates but diurnal fluctuations can be quite large, especially during a shift in wind direction from the north-west to the south. Strong winds and heavy precipitation are often associated with north-westerly winds which occur most frequently in the late winter and early spring.

The data used in his study were collected at the Ski Basin climate station (elevation 1500 m a.s.l.) located on a 2° slope at the base of a south-east facing cirque (Fig. 2). It is the only daily-serviced, year-round climate station above the winter snowline in the South Island.

Energy Balance

For an isothermal snowpack of 0°C, the energy balance may be written (Male and Gray, 1981) as:

$$Q_m = Q^* + Q_h + Q_e + Q_p + Q_g \qquad (1)$$

where Q_m = heat available for melt (W/m^2)
Q^* = radiation heat flow (W/m^2)
Q_h = sensible heat flow (W/m^2)
Q_e = latent heat flow (W/m^2)
Q_p = precipitation heat flow (W/m^2)
Q_g = ground heat flow (W/m^2)

In this and subsequent energy transfer equations, positive signs indicate energy gains to the snowpack. These energy transfers are given as flux densities (rates of flow per unit area). However, the term 'flow' is used in this sense throughout the remainder of the paper.

Radiation Heat Flow

Radiation heat flow is expressed as the balance of short- and long-wave radiation (Geiger, 1966) and has traditionally been expressed as:

$$Q^* = S^* + L^* = S\downarrow(1-a) + (L\downarrow - L\uparrow) \qquad (2)$$

where S^* = net short-wave radiation (W/m^2)
L^* = net long-wave radiation (W/m^2)
$S\downarrow$ = direct and diffuse solar radiation (W/m^2)
a = snow-surface albedo to shortwave radiation
$L\downarrow$ = atmospheric long-wave radiation (W/m^2)
$L\uparrow$ = terrestrial long-wave radiation (W/m^2)

For this study, incoming short-wave radiation was measured with a Feuss actinograph but no direct measurement of the surface albedo was taken. An albedo of 0.5 was assumed for melting snow (U.S.A.C.E., 1956).

Long-wave terrestrial radiation was computed as a function of surface snow temperature and snow emissivity (assumed to be 0.99, Anderson, 1976). The Brutsaert (1975) equation, adjusted for alpine use (Marks, 1979), was used to predict atmospheric radiation. It was modified also to account for the degree and type of cloud cover (Sellers, 1965) and for the view factor (Lee, 1962) of the study site.

The complete long-wave radiation balance was therefore expressed as:

$$L^* = 1.24(e'_a/T_a')^{0.14}\,(P_a/1013)\sigma T_a^4\,(1+\Omega\phi^2)\cos^2(90-H) + \varepsilon_s\sigma T_s^4(1-(\cos^2(90-H))) - \varepsilon_s\sigma T_s^4 \quad (3)$$

where e_a' = near-surface vapour pressure (mb)
T_a = near-surface air temperature (°K)
T_a' = near-surface temperature (°K) adjusted to sea level
P_a = near-surface air pressure (mb)
σ = the Stefan-Boltzman constant (5.6697×10^{-8} W m^{-2} K^{-4})
Ω = cloud height and temperature coefficient
ϕ = the fraction of sky obscured by cloud
H = average horizon angle from the zenith (degrees)
ε_s = snow emissivity
T_s = snow-surface temperature (°K)

Measurements of air temperature were obtained from thermograph records and surface snow-temperatures were assumed to be at the melting point during melting. Vapour pressure was estimated from hygrograph records according to Rose's (1966) method. Cloud cover was obtained from daily observations.

Sensible and Latent Heat Flow

The fluxes of sensible and latent heat flow were calculated according to the following equations which have been reviewed by Kuzmin (1961), Anderson (1976) and Male and Granger (1979).

$$Q_h = \rho_a C_p D_h (T_a - T_s) \quad (4)$$

$$Q_e = \rho_a L_v D_e (0.622/P_a)(e_a - e_s) \quad (5)$$

where ρ_a = air density (kg/m^3)
C_p = specific heat of air at constant pressure (J kg^{-1} K^{-1})
L_v = latent heat of vapourisation of water (J/kg)
D_h = exchange coefficient for sensible heat (m/s)
D_e = exchange coefficient for latent heat (m/s)

The snow-surface vapour pressure was obtained by assuming that the air in contact with the snow was saturated. Values of the saturated vapour pressure over ice were obtained from standard meteorological tables (List, 1966).

The values of the exchange coefficients for sensible and latent heat were assumed to be equal to that for momentum as described by many researchers (Sellers, 1965; Priestley, 1959; Anderson, 1976) and were expressed as:

$$D_h = D_e = D_m = k^2u/[\ln(z/z_o)]^2 \quad (6)$$

where k = von Karman's constant
u = wind velocity (m/s) at height z (m)
z_o = a roughness parameter (m)

Von Karman's constant was set to 0.4 (Businger, 1973) and surface roughness at 2.5×10^{-3} m (Sverdrup, 1936). Stratified conditions (lapse or inversion) were adjusted according to the method of Price and Dunne (1976):

$$D_u = D_m/(1 + \alpha Ri) \quad (7)$$
$$D_s = D_m(1 - \alpha Ri) \quad (8)$$

where D_u = exchange coefficient under unstable conditions (m/s)
D_s = exchange coefficient under stable conditions (m/s)
α = a constant approximately equal to 10 (Webb, 1970)
Ri = the bulk Richardson number expressed as:

$$Ri = gz\Delta T_a/[T_a(\Delta u)^2] \quad (9)$$

where g = acceleration due to gravity (m/s^2)
ΔT = temperature difference (K) over the height z (m)
Δu = windspeed difference (m/s) over the height z (m).

These formulations assume steady, turbulent flow over an infinite, uniform surface, conditions which are certainly not met in mountain environments. Male and Granger (1979, p. 118) show that these equations result in considerable errors even when corrected for stability. Consequently, the results of using this approach should be considered as first approximations.

Precipitation Heat Flow

Rain may also contribute heat to a melting snowpack, the total flux expressed as (Male and Granger, 1979):

$$Q_p = \rho_w C_w P_i (T_p - T_s) \times 10^{-3} \quad (10)$$

where ρ_w = density of water (kg/m^3)
C_w = specific heat of water (J kg^{-1} K^{-1})
P_i = precipitation rate (mm/s)
T_p = precipitation temperature (°K)

Precipitation was measured with a Belfort weighing-bucket precipitation gauge, and the wet-bulb air temperature was used as a surrogate for rainfall temperature as suggested by Anderson (1976).

Ground Heat Flow

The remaining heat flux in equation 1 is ground heat flow and for most short-term snowmelt events is treated as negligible relative to the other heat flow terms, although over the long term it may supply significant quantities of heat to the snowpack. Ground heat flow at the study site was estimated from measured soil temperature gradients and approximate thermal conductivities of the soil. The total ground to snow flux never execeeded 0.65 MJ m^{-2} d^{-1}, a value consistent with results reported elsewhere (for example: Yoshida, 1962; Granger, 1977; Gold, 1958 and Kuzmin, 1961). Soil surface temperatures are often at or below the freezing point during spring snowmelt, and the ground may even act as a heat sink. This small amount of ground heat flow is not included in further discussion or calculation of the heat flows in the remainder of this paper.

Heat flows were calculated using 24 h totals of short-wave radiation and windrun, and daily means of air temperature and humidity.

SELECTION OF MELT PERIODS

Melt periods were selected from a record of daily snow depth measured at the Ski Basin climate station over the years 1975 to 1980 (Fig. 3). A

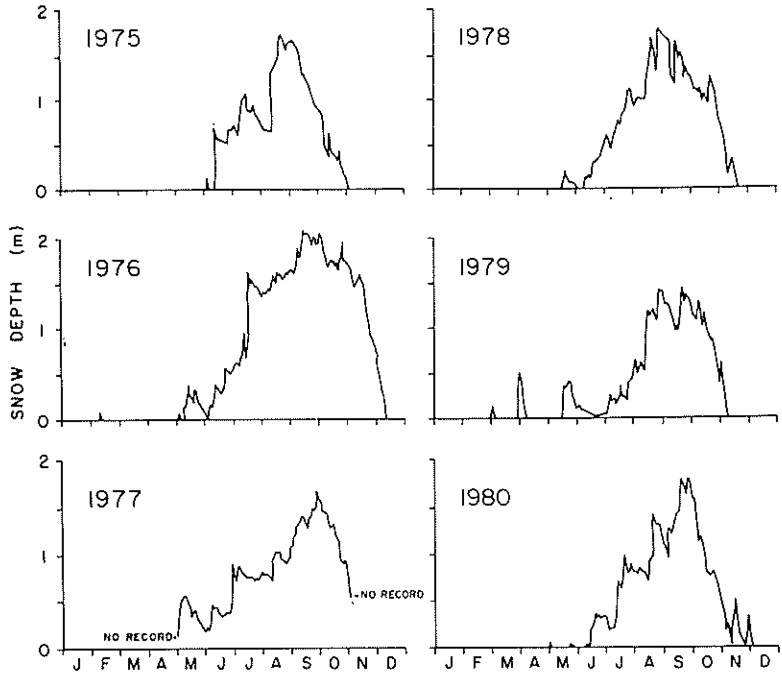

FIG. 3—Daily snow depth measured at Ski Basin climate station.

number of problems exist in using simple snow depth to indicate snowmelt. Changes in depth may reflect wind deflation and erosion, and settlement with metamorphism. Snow depth alone does not provide information on changes in the water equivalent of a snowpack. Snow depths cannot be used for estimating snowmelt without information on snow density.

Because of these problems, only periods of rapid snow depth depletion, covering several days and preceded by a snowfall-free period of relatively constant snow depth, were analyzed to ensure that the depth decreases were not simply due to settlement following fresh snowfalls. Because temperatures in the Craigieburn Range are relatively warm, most of settlement from metamorphism occurs in the few days immediately following a snowstorm. Settlement rates are only slight after this phase and subsequent rapid decreases in depth are probably due to melt.

The selected periods were characterized by near-surface air temperatures well above the freezing point, which also is an indication of likely melting. Since the surface snowpack was likely to be melting, the surface adhesion would be relatively high, reducing the possibility of wind erosion. The location of snow depth measurements is also believed not to be within a zone of noticeable wind deflation. Four spring melt periods were

TABLE 1—Weather conditions during spring snowmelt.

Period	Air temp. (°C)	Max air temp. (°C)	Windspeed (m/s)	Global radiation ($MJ\ m^{-2}\ d^{-1}$)	Vapour pressure (mb)
Nov. 15-28, 1976	3.8(−1.1)	7.4(−1.2)	2.7(−0.1)	16.50(−3.54)	6.36
Oct. 26-30, 1977	7.3(+4.8)	10.8(+4.9)	4.5(+1.5)	20.40(+3.17)	5.76
Oct. 24-29, 1979	6.0(+3.5)	8.9(+3.0)	4.2(+1.2)	17.59(+0.36)	6.78
Oct. 22-29, 1980	7.4(+4.9)	10.8(+4.9)	3.4(+0.4)	16.48(−0.75)	6.61

Values given are means for each study period. Figures in parentheses show deviation from long term monthly means.

selected for study, ranging in duration from five to fourteen days (Table 1). In general, the periods were characterized by either one or more series of clear skies followed by increasing air temperatures and cloud. For the three October periods, the mean windspeed and air temperatures were above the long-term mean monthly average, while in November both parameters were below normal. Extensive cloud in November reduced the levels of global radiation to less than or equal to those recorded in the three October periods. The greatest rate of radiation flux was received between October 26 and 30, 1977, when global radiation was 20 $MJ\ m^{-2}\ d^{-1}$, 18 percent in excess of the long-term mean monthly value.

The diversity of weather conditions which characterize the melt periods is important. Much of the early energy balance work focused on periods with relatively constant meteorological conditions, whereas, more recently, researchers (Granger and Male, 1978; McKay and Thurtell, 1978) have pointed to the importance of changing meterological conditions in establishing the long-term melt characteristics within a region.

RESULTS AND DISCUSSION

The calculated daily heat flows for the four melt periods appear in Figure 4, and the relative percentage contribution of each heat flow to the total heat input are listed in Table 2. Sensible heat was the major heat component on every day except one during the three October melt periods, contributing 60 to 64%. For the fourteen-day period in November, sensible heat flow and radiation heat flow dominated the heat flow on seven days, each contributing approximately the same percentage (42 and 43%) to the total heat input. The greater importance of radiation heat flow in November compared to the October periods might normally be expected to be due to the seasonal increase in solar radiation. However, as mentioned earlier, the mean daily radiation receipts in November were as low or lower than those recorded in the October periods. The main reason for the increase in the importance of radiation heat flow in November is the decrease in the sensible heat supply. Low mean windspeeds and air temperatures during November (Table 1) were responsible for the low sensible heat transfer.

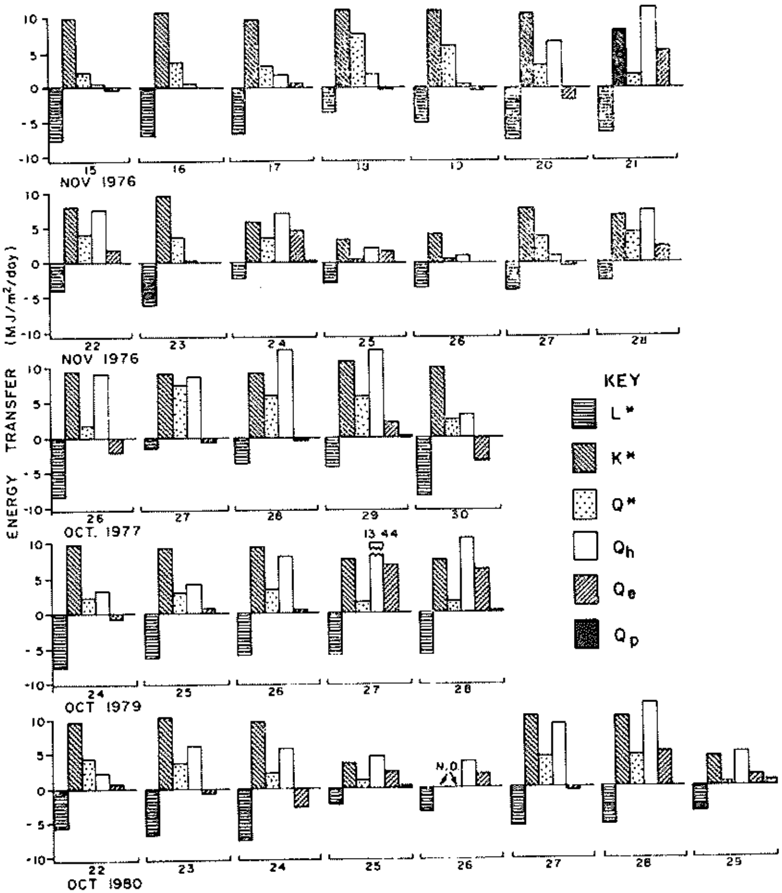

FIG. 4—Magnitude of major heat flows to the snowpack. N.D. refers to missing data.

TABLE 2—Percentage contribution total heat supply to snowpack.

Period	Q*	Q_h	Q_e	Q_p
Nov. 15-28, 1976	42	43	14	<1
Oct. 26-30, 1977	33	64	3	<1
Oct. 24-29, 1979	17	61	21	<1
Oct. 22-29, 1980	27	60	12	1
Average	30	57	13	<1

TABLE 3—Percentage of total heat input on days of greatest heat supply.

Date	Q^*	Q_h	Q_e	Q_p
Nov. 21, 1976	9	62	29	<1
Oct. 29, 1977	29	59	10	2
Oct. 27, 1979	7	62	31	<1
Oct. 28, 1980	21	56	23	<1

TABLE 4—Percentage of heat supply and precipitation characteristics on days with rain.

Date	Q^*	Q_h	Q_e	Q_p	Precip (mm/d)	Temp (°C)
Nov. 24, 1976	23	46	30	1	11.5	3.4
Nov. 25, 1976	10	48	40	2	6.8	2.4
Oct. 29, 1977	29	59	10	2	19.0	5.5
Oct. 29, 1980	4	68	20	8	22.0	4.4

(Precipitation temperature derived from wet-bulb air temperature).

In total, sensible heat produced the greatest heat supply to the snowpack on 23 of the 31 days analyzed (radiation data was not available for October 26, 1980). The dominating importance of sensible heat contrasts with the results of Harding (1972) and Anderton and Chinn (1978) who found that, over melting snow and ice surfaces, net radiation accounted for most of the heat supply. However, Anderton and Chinn (1978) note that their work was conducted under relatively calm, clear-sky conditions during summer. Radiation receipts would naturally be high during such periods, and low windspeeds would reduce the sensible heat flux.

Latent heat also was important in snowmelt. Rapid latent heat transfer is usually associated with moist warm air-masses which develop a strong vapour pressure gradient decreasing towards the snow surface. This condition often prevails during north-westerly storms which frequently invade the Craigieburn Range, particularly in late winter and early spring. One such storm developed over the melt period October 24-29, 1979, when latent heat flow contributed over twenty percent of the total heat input, and exceeded 5 MJ m^{-2} d^{-1} on October 27 and 28. On many of the days studied, the contribution of energy from the latent heat of condensation even exceeded that from net radiation, but in all cases remained below that from sensible heat flux.

Overall, precipitation heat flow was relatively unimportant, contributing less than one percent of the total heat supply in all periods (Table 2). The days which recorded the greatest total heat supply within each of the four melt periods, however, were all characterized by rain. Table 3 lists the relative importance of the heat flows on these days. Similar to the averages for the entire melt periods (Table 2), sensible heat remained the dominant flux, contributing approximately sixty percent (56 to 62) of tne total heat supply. As might be expected during rain, the contribution from net radiation was appreciably reduced and the flow of latent heat increased to exceed the radiation flux on these days.

The relatively small heat contribution from rain contrasts with the finding of Fitznarris *et al.* (1980) that precipitation heat flow contributed up to twenty percent of the total heat input during a rain-on-snow event in southern Otago, and of Anderton and Chinn (1978) that rainfall provided 19 and 11 percent of the total heat supply on two successive days of rain.

To further assess the role of precipitation heat flow in this study, all days which recorded more than 5 mm of rain were analyzed for the relative importance of the heat flows (Table 4). In general, the relative importance of radiation, sensible and latent heat flows remained approximately the same as on the days of maximum heat input (Table 3). Minor percentage decreases in sensible and latent heat flows were taken up by precipitation heat flow which averaged only 3 percent and never exceeded 8 percent of the total daily heat supply.

The main reasons for the difference in the importance of precipitation heat flow in this study and in those by Fitzharris *et al.* (1980) and Anderton and Chinn (1978) are the magnitude and temperature of the rain. Fitzharris *et al.* (1980), in particular studied exceptionally intense rain (10 mmh for 25 hours) and rain temperatures of 10°C. In contrast, the amounts of rain were smaller and the temperatures cooler in this study. The maximum daily rainfall was only 22 mm and the maximum rain temperature only 5.5°C. These conditions are probably more typical of most rain-on-snow events for the Craigieburn Range; those reported by Fitzharris *et al.* (1980) would be representative of a rare event.

Net radiation is usually the parameter controlling snowmelt when air masses are well established, especially under calm conditions with clear skies (McKay and Thurtell, 1978). Sensible heat frequently becomes the most important heat flow when warm air is advected into a region. Because of the overall importance of sensible heat identified in this study, the air mass directions under which large fluxes of sensible heat occur were analyzed. Strong wind and warm air are both required to raise sensible heat flow. Days were selected from the period May to November over a five year period, 1975 to 1979, if they recorded:

a) a mean temperature greater than the long-term mean monthly maximum temperature, and

b) an average daily windspeed equal to or greater than 5 m/s.

Seventy-four such days were identified (Table 5). The magnitudes of the sensible heat flows are very similar to those reported for environments where sensible heat is known to be important for snowmelt. For example, Golding (1978) reports that during winter Chinooks in Canada, sensible heat contributed on average to the snowpack 3.4 MJ m^{-2} d^{-1} in 1975 and 5.7 MJ m^{-2} d^{-1} in 1976. For a comparable period (July, August and September) in the Craigieburn Range, the mean input is 5.8 MJ m^{-2} d^{-1}

TABLE 5—Mean daily sensible heat transfer for days with a mean temperature greater than the long-term monthly maximum and with a mean windspeed >5 m/s.

Month	Sensible heat transfer ($MJ\ m^{-2}\ d^{-1}$)*	Number of days
May	10 ± 4	8
June	6 ± 3	7
July	3 ± 2	11
August	7 ± 6	11
September	7 ± 4	16
October	12 ± 6	15
November	12 ± 3	6

* Mean values and standard deviations.

(Table 5). Values in excess of 11 $MJ^{-2}\,d^{-1}$ are reached in October and November.

The directions of air flow on days of high sensible heat transfer were obtained from radiosonde records (700 to 800 mb pressure level) for Christchurch Airport (Fig. 5). For over 65% of the 74 days, the dominant airflow was from the northwest. The elevation of freezing levels in the Craigieburn Range is at a maximum during north-westerly airflows and hence the highest air temperatures also occur under such wind directions (Prowse, 1981). These are precisely the conditions which produce high sensible heat flow.

Although the Craigieburn Range is in the rainshadow of the Southern Alps to the west, north-westerly storms there are usually characterized by high humidity and precipitation. Some storms, however, combine strong winds with low humidities because of a föhn effect and sublimation of snow may become important to total ablation. Further research is needed to quantify the long-term characteristics of north-westerly storms, and their role in snowmelt production.

CONCLUSIONS

During spring snowmelt, sensible heat is the major source of energy to the snowpack, contributing approximately sixty percent of the total heat supply. Above average air temperatures and windspeeds, typical of north-westerly airstreams, are primarily responsible for the dominating influence of the sensible heat flux. Over the long term, daily inputs of sensible heat under such north-westerly conditions are similar in magnitude to those reported for Chinooks in Canada.

Net radiation is the second most important heat flux, reaching maximum levels during calm clear periods late in spring. Under heavy cloud, high humidity and strong wind, radiation heat flow is less than latent heat flow. The greatest total heat flux to the snow pack occurs on days with rain, although the precipitation heat flow is relatively unimportant, less

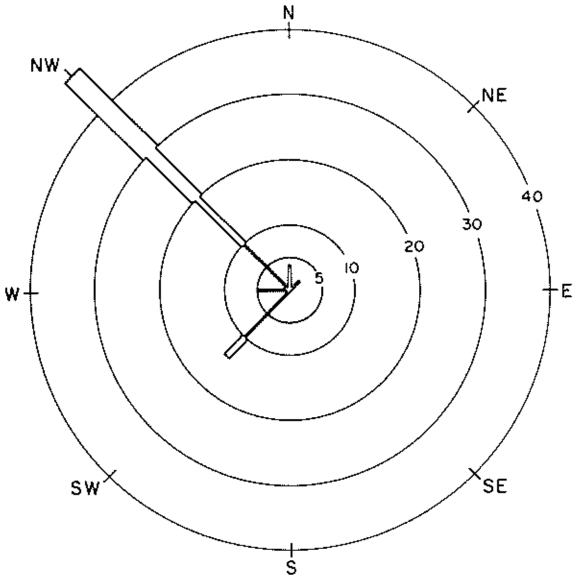

FIG. 5—Frequency of upper air wind direction on 74 days of high sensible heat transfer.

than eight percent of the total heat supply. Sensible and latent heat are the most important heat supplies during rain.

Future studies of snowmelt should concentrate upon obtaining more data on the size and relative importance of heat flows during specific weather systems and especially during wet north-westerly airflows. Attempts should also be made to identify the long-term frequency of such events and hence the overall importance of individual heat flows to total snowcover ablation. Such information is important in selecting techniques for modelling and predicting snowmelt. Attention should also be paid to sublimation, which under föhn conditions may be siginficant.

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