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Assessing snow storage and melt in a New Zealand mountain environment

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ABSTRACT There are no systematic measurements of the snow resource in New Zealand, so it has to be inferred from records from lowland stations. In this way, the water-balance method is used to compute seasonal snow storage for a grassland, wind-swept, mountain basin of area 120 km^2. The energy sources for melt are assessed for a major flood. Runoff is maximum in October, when it averages 2.8 times the mean annual flow. Mean snow storage over 10 years was 176 mm, or one third of annual flow. During the flood, a snow pillow recorded 97 mm of melt in 43 h. It is calculated that over half the energy at the melting snow surface came from convection of sensible and latent heat.

INTRODUCTION

Snow hydrology is in its infancy in New Zealand, and information on snow is seldom used in water-manangement decision making. There are no regular, quantitative estimates of seasonal snow storage, even though the country is mountainous, and over three quarters of its electricity supply is generated by hydro power. A few snow courses have been established in the past (e.g. Gillies, 1964; Chinn, 1969; Archer, 1970) but these are no longer systematically maintained. All but a few climate stations are below the winter snowline. As a consequence, the size of the snow storage has to be estimated from analyses of runoff and meteorological records from low-altitude stations.

The objective of this paper is to calculate the size of seasonal snow storage for the Fraser basin, in the South Island of New Zealand, by using a water-balance approach. Snowmelt, and its sources of energy, are assessed for a major flood (11-17 October 1978), which Jowett (1979) estimates to be the largest for a century.

The Fraser basin is chosen because it is sufficiently high to be largely snow covered in winter and probably represents an upper limit to the relative size of seasonal snow storage expected in New Zealand basins. Its runoff is used for both hydro power generation and for irrigation. Gillies (1964) considers this IHD Representative Basin to be typical of some hundreds of kilometres of surrounding range country (see New Zealand.Ministry of Works 1970).

NATURE OF THE FRASER BASIN

The Fraser basin at latitude 45°S has an area of 120 km^2, and varies in elevation from 548 m to 1695 m, with a mean elevation of 1400 m. Almost 90% of the area is above 1000 m (Fig.1), the average seasonal snowline. Grimmond (1980) calculates mean flow for 1969-1978 as 1.9 m^3s^{-1}. The gauging site has a stepped concrete pipe weir, where the water level is measured with a Fischer and Porter recorder.

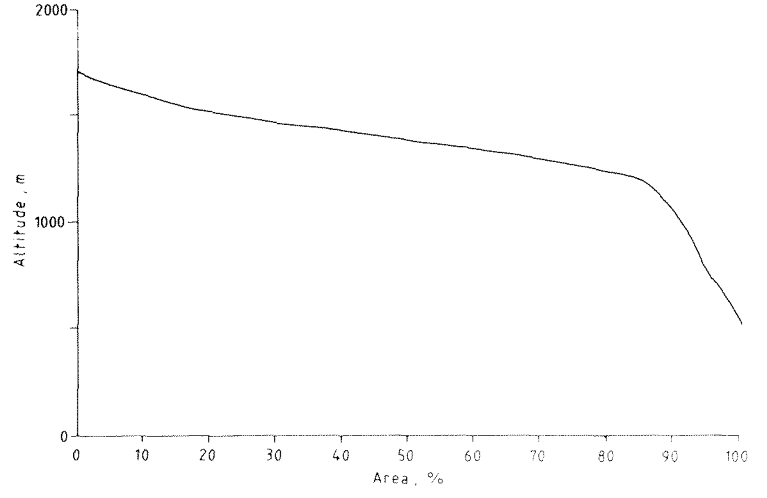

FIG.1 Hypsometric curve for the Fraser basin.

The terrain is gently rolling, with over two-thirds having a slope of less than 15°. Broad interfluves are dissected by small, but entrenched gullies. Coarsely foliated schist underlies the basin, so that it is considered to be watertight (Harrison, 1978). At lower altitudes, vegetation consists of tussock grassland, 0.2-0.5m tall. At higher altitudes, the grassland grades into shorter alpine herbs and cushion fields. The tussock grasses act as traps for blowing snow and play an important role in determining runoff and the timing of melt (Fitzharris, 1976; Harrison, 1978).

Strong winds are common, with average monthly speeds of 6-11 m s^{-1} (Bliss & Mark, 1974), so that large snow drifts form around solid objects, in gullies, and in the lee of ridge crests. Harrison (1978) shows that aspect, slope and breaks in topography are major determinants of snow accumulation within the basin. On the broad interfluves the winds restrict accumulation to depths of less than one metre, but snow can exceed 10 m in gullies, where small swamps also occur.

Gillies (1964) notes that the snow cover does not accumulate as a continuous process, but that it thaws, recedes and reforms at intervals throughout the winter in response to wide and frequent fluctuations in atmospheric freezing level. Thus it is difficult to recognize clearly defined accumulation and melt seasons. In most years, all snow melts by late summer.

METHODS

The water balance for the basin is estimated, using monthly means for the period 1969-1978, as follows

$$P - R - E \pm \Delta S = 0 \tag{1}$$

where P is the precipitation, R the runoff, E the net evaporation-condensation and ΔS the change in storage, here assumed to be of snow, soil water (including swamps), and groundwater.

Basin runoff records provide values of R. E is calculated using the method of Priestly & Taylor (1972):

$$E = \alpha\ S\ Q_R/L_v \tag{2}$$

where α is a dimensionless coefficient, which Brash & Murray (1980) found to be unity in a nearby basin (the same value is assumed here), L_v is the latent heat of vaporization,

$$S = \Delta\ /(\Delta + \gamma)$$

where Δ is the slope of the saturated vapour pressure curve at the wet bulb temperature and γ is the psychrometric constant. Q_R the net radiation, is given by:

$$Q_R = G\ (1 - a) + L^* \tag{3}$$

where G is the global radiation, a the albedo of the basin and L* the net longwave radiation.

G is assumed to be that at Alexandra (elevation 141 m) 10 km to the east, where measurements are made with a Robitsch type actinograph. Albedo varies throughout the year according to the proportion of the basin covered by snow. For snow-free conditions, Brown (1977) measured a = 0.24 over tussock. The method of Idso & Jackson (1969) calculates L* for clear sky conditions, and values are adjusted for an average cloud-cover fraction of 0.7, using the procedures in Oke (1978). Temperatures required for these and subsequent calculations are extrapolated from valley climate stations at Alexandra or Roxburgh (10 km to the east, elevation 110 m), to the mid-elevation of the basin using a lapse rate of 0.88 deg/100 m (Grimmond, 1980).

Estimates of mean annual P are obtained from Jowett & Thompson (1977), and adjusted using the methods of Hare & Hay (1971) and Brash & Murray (1980). The overriding assumptions are that errors in the estimation of mean energy-balance components for an area, particularly Q_R and L_vE, are less than those associated with mean precipitation based on sparse point measurements in a windy environment, and that over the long term $\Delta S = 0$. From this procedure, annual P = 869 mm, or 2.55 times the precipitation at Alexandra. This relationship is used to estimate monthly values of P for the basin.

A snow pillow (diameter 1.8 m) recorded melt at an elevation of 1370 m during the flood. The energy balance for the melting snow surface is

$$Q_R + Q_S + Q_L + Q_M + Q_r = 0 \tag{4}$$

where the subscripts S, L, M and r refer to sensible heat flux, latent heat flux of evaporation, latent heat flux of melting, and melt supplied by rain, respectively. Fluxes directed towards the surface are positive.

Again, Q_R records are obtained from Alexandra. $(Q_S + Q_L)$ are not calculated, but found as residuals in equation (4). $Q_M = L_f M$, where M is the melt as indicated on the snow pillow record, and L_f the latent heat of fusion.

$$Q_r = Cr(T_r - T_s) \tag{5}$$

where C is the specific heat of water, r the rainfall rate, T_r the rain temperature and T_s the snow surface temperature (= $0^{o}C$).

RESULTS

Runoff rises in September, to peak in October, with high flows sometimes extending into January. The increase occurs as a bulge in baseflow, on which are superimposed clearly defined peaks (see the runoff record for 1974 given in Fig.2 as an example). These

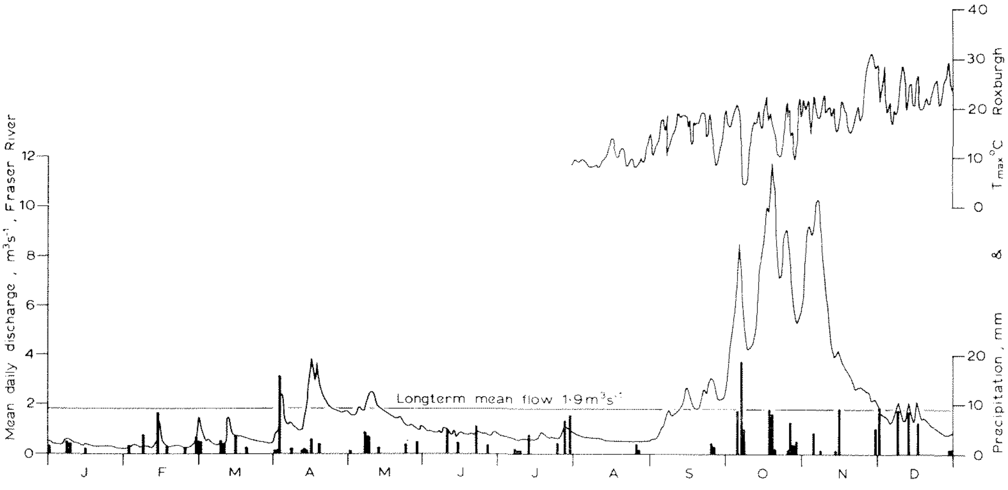

FIG.2 Runoff pattern for the Fraser basin for 1974. Daily precipitation and maximum temperature are shown for Roxburgh.

features are produced by snowmelt as temperatures rise in the spring, and by rain or snow which may occur at any time. River flow declines in late summer as the snow disappears. A secondary flow maximum occurs in autumn due to rainfall from occasional subtropical depressions and a reduction in evaporation. Winter is the time of lowest flow because of water entering into snow storage.

October flow is 2.8 times the annual mean, while that in July is reduced to 0.4 times the mean.

On average, ΔS and accumulation ($\Sigma\Delta S$) vary systematically throughout the year (Table 1). There are net storage gains from

TABLE 1 Water balance for Fraser basin (1969-1978) and values of albedo (averaged over basin)

Month	*P (mm)*	*R (mm)*	*E (mm)*	*ΔS (mm)*	*$\Sigma\Delta S$ (mm)*	*a*
Jan	70	29	75	- 34	22	0.24
Feb	51	16	57	- 22	0	0.24
Mar	81	19	39	23	23	0.24
Apr	75	26	8	41	64	0.24
May	85	33	0	52	116	0.30
Jun	57	30	0	27	143	0.40
Jul	55	16	0	39	182	0.70
Aug	60	21	0	39	221	0.70
Sept	83	53	0	30	251	0.70
Oct	102	124	29	- 51	200	0.40
Nov	47	104	59	-116	84	0.30
Dec	103	57	75	- 28	56	0.24
Annual	869	528	341	0	-	0.39

March to September, and net losses from October to February. Snow accumulation is obtained by subtracting the other storages: changes in groundwater storage are neglected; substantial seasonal changes in soil moisture have been observed below 900 m, but diminish with elevation to become small above 1400 m (Mark, 1965). To allow for these, and changes in water level of small swamps associated with snow-patch gullies, soil storage change is estimated at 75 mm. On this basis, net snow accumulation begins in May and reaches a peak of 176 mm in September, which is equivalent to 33% of the mean annual flow. Net storage losses begin in October and are most rapid in November (-116 mm). Some represents evaporation losses from the soil, but a substantial portion is snowmelt.

Over individual years maximum snow accumulation reached 276 mm in August 1976, but only 80 mm in July 1977. This suggests a 10 year variation about long term mean storage of ±57%.

During the flood, rainfall totalled 80-150 mm within the Fraser basin, with 150 mm at the snow pillow, mainly on 13-14 October 1978. Peak flow at the gauging site is estimated at 88-106 $m^3 s^{-1}$ (Fitzharris *et al.*, 1980).

The snow pillow indicated rapid melt over 43 h on 12-14 October, with a net loss of 97 mm water-equivalent at this site (Fig.3). Melt averaged 2-3 mm h^{-1}, with 40% occurring before significant rain during gusty northwest winds and temperatures that rose to 20^{o}C at Roxburgh. The remaining 60% of melt was accompanied by rain and cooler temperatures of 5-10^{o}C at Roxburgh. At the snow pillow, the average values of the energy sources for melt over the 43 h are

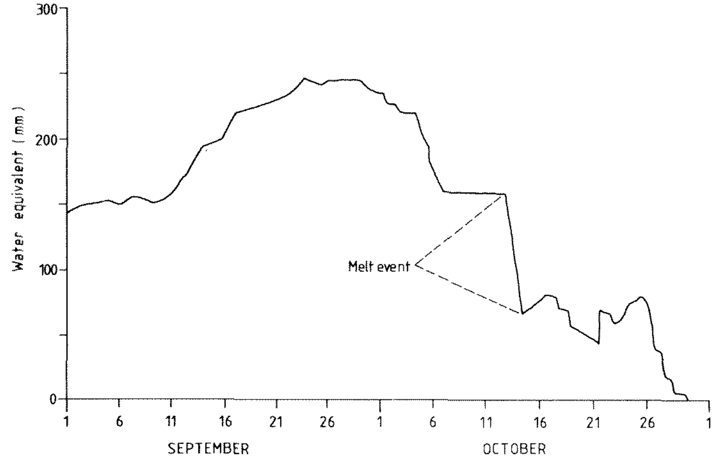

FIG.3 Snow-pillow record showing the major melt event during the 1978 flood.

calculated as: Q_R = 47 W m^{-2} ; Q_r = 41 W m^{-2} ; Q_M = -209 W m^{-2} ; and (Q_S + Q_E) = 121 W m^{-2} . Thus the convective fluxes provided over half of the energy for melt. Presumed gradients of temperature and humidity above the snow surface favour the dominance of Q_S in the dry conditions before the rain and Q_L after its commencement (Fitzharris *et al.*, 1980). Over the Fraser basin, storm rainfall averaged 135 mm and, after taking into account the elevation of the snowline, snowmelt was estimated at 80 mm. Thus melt contributed 37% of the total water input during the flood.

DISCUSSION

Seasonal snow storage that is 33% of annual mean flow is the highest reported for a New Zealand basin. Anderton (1976) estimated that snowmelt and ice-melt contributed 21% of runoff from the Ivory basin (elevation range 1400-1700 m) during 1971-1975. He also considered that snowmelt supplied 20% of the mean annual inflow to Lake Pukaki (480-3764 m) (Anderton, 1974). For the Clutha basin (130-3036 m), Jowett & Thompson (1977) indicate that average maximum accumulation of snow is 255 mm, or 14% of runoff. In his review of the issue, Fitzharris (1979) concluded that for large South Island basins, snow storage supplies 10-25% of annual flow.

October flow that is 2.8 times the annual average is similar to that given by Ward (1978) for mountain spring snowmelt regimes of the Northern Hemisphere. He uses as examples the rivers Fraser (Canada), Reusz (Switzerland) and Inn (West Germany).

Kuz'min (1961) notes that melting rates of 9 mm h^{-1} have been measured in the European USSR, and that rates of 3 mm h^{-1}, similar to those observed here, are common in warm, humid, windy conditions. Mild, moist northwest airflows are a feature of New Zealand spring weather, suggesting that rapid melt from convection of sensible and latent heat toward the snow surface may be a common occurrence. Similar, but lesser events are observed in other years (see early

October 1974 in Fig.1). In the case of the major October 1978 flood, melt contributed to a rapid rise of the river and effectively wetted the basin, so priming it for rapid runoff of the subsequent heavy rain.

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