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A CONCEPTUAL RUNOFF MODEL FOR A MOUNTAINOUS RAIN-ON-SNOW ENVIRONMENT, CRAIGIEBURN RANGE, NEW ZEALAND

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ABSTRACT

A runoff model, incorporating a distributed snow routine and a lumped transformation routine, was developed for conditions in non-glacierized mountainous basins in the eastern South Island of New Zealand. The model, which is based upon the Swedish HBV model, was applied to the Camp Stream catchment in the Craigieburn Range. A major problem was over-estimation of runoff, which made calibration of the runoff parameters difficult. Mean daily air temperature appears to be inadequate for discriminating rain from snowfall at elevations lower than about 1500 m. A snowmelt routine incorporating an index of the strength of regional air flow reproduced the variability of runoff during a snowmelt better than did a simpler degree-day melt model. The lumped transformation routine is inadequate when runoff sources and mechanisms vary between events. Further modelling work on this environment should include distributed transformation routines and alternative wind-speed indices for snowmelt calculations.

INTRODUCTION

Conceptual models of snowmelt and rainfall runoff have been used for many purposes, including water-supply forecasting and flood prediction. Conceptual parametric models applied in New Zealand have so far ignored snow accumulation and melt processes, although many of the large South Island rivers which drain mountain catchments have significant snowmelt (Fitzharris, 1979). This paper presents a model of snow accumulation, melt and runoff which can be used for the South Island of New Zealand, an area of rugged terrain, mid-winter melt and frequent rain-on-snow events. Conceptual models, which explicitly represent many of a system's unit processes, were chosen because they have been used successfully to model snowmelt runoff (Anderson, 1979). The Swedish HBV-3 model (Bergstrom and Johnson 1976) was selected because of its simple structure, modest input requirements, inclusion of a satisfactory snow routine and successful application in a number of environments.

STUDY AREA AND DATA BASE

The modelling exercise was carried out using data collected on the east side of the Craigieburn Range in the Waimakariri River catchment, approximately 20 km east of the Main Divide of the South Island (Fig. 1).

The Craigieburn Range is characterized by rounded ridges, and valley side slopes of 30 to 40° above 1200 m, with many bedrock outcrops on the upper slopes. Evergreen forests composed of several species of beech (*Nothofagus* spp.) extend up to 1370 m, although extensive areas of forest have been cleared. Scree and snow tussock grassland dominate areas above treeline.

Mean annual precipitation measured at various sites ranges from 1500 to 2000 mm. Precipitation is well distributed throughout the year, with peaks in autumn and spring. Snow may fall at any time of year, but is usually

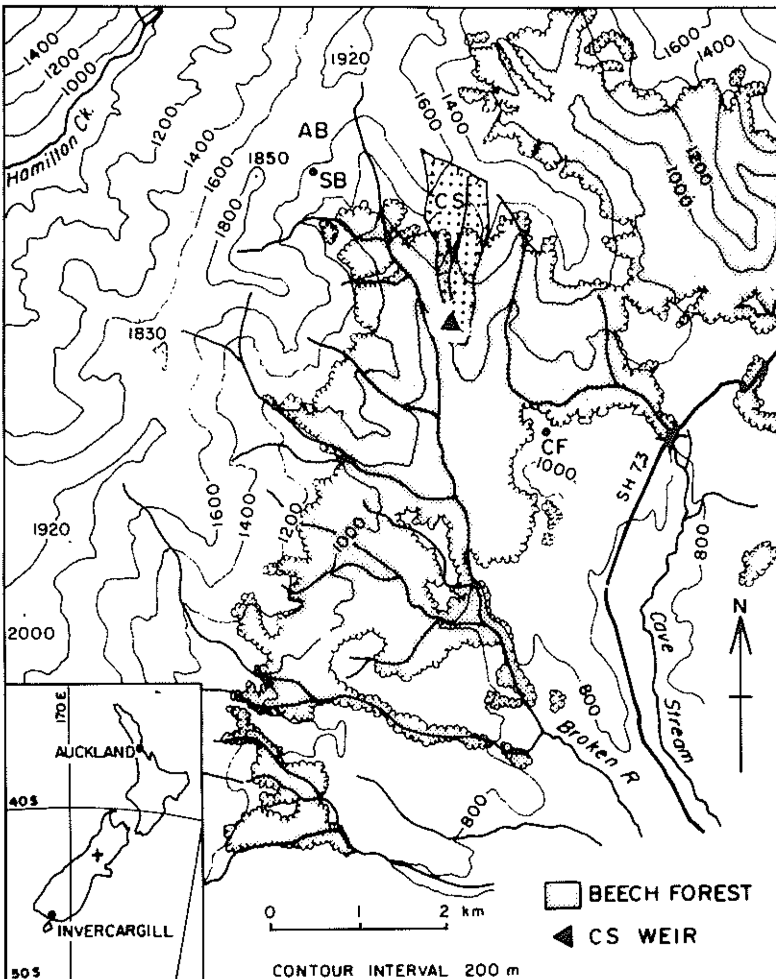

FIG. 1: Location map
AB denotes Alan's Basin
SB denotes Ski Basin
CS denotes Camp Stream
CF denotes Craigieburn Forest

concentrated in the winter months. Snow is roughly one-third of annual precipitation above 1500 m, but is negligible below 900 m (O'Loughlin, 1969).

Snow accumulation varies substantially in water equivalent from year to year. The length of the snow season varies, but snow usually begins accumulating in May or June and melts by late November (O'Loughlin, 1969). The mean seasonal snow-line lies at approximately 1300 m, but snow often falls down to 700 m and subsequently melts several times in a winter. Even above seasonal snow-line, snow accumulation is interrupted by occasional mid-winter rain and melt, the frequency and magnitude of which decrease with elevation. The bulk of seasonal snow accumulation melts between September and November, when the frequency of north-westerly winds and rain is greatest.

The New Zealand Forest Service and the Ministry of Works and Development (MWD) have maintained data collection programmes in the Craigieburn Range since the 1960's. These include climate stations, snow surveys and streamflow gauging sites (O'Loughlin, 1969; Ministry of Works, 1968). Snow surveys in Alan's Basin and Camp Stream were discontinued after 1973. Only the Ski Basin and Craigieburn Forest climate stations (Fig. 1) have been visited daily, and Ski Basin was visited only weekly after November 1983. A recording weighing-bucket precipitation gauge is presently maintained by MWD at 1370 m elevation in Camp Stream catchment. A gauging site is maintained on Camp Stream (Fig. 1), although the quality of the record is variable. Camp Stream is a first-order catchment with an area of .94 km^2 and an elevation range of 1020 to 1730 m. The distribution of vegetation and hypsography are shown in Table 1.

Daily observations of maximum and minimum temperatures and precipitation are the most readily available climate data for South Island mountain areas. It was therefore decided initially to constrain the model to require only temperature and precipitation data and other readily-available information, even though wind speed and global radiation are available for one or both of Craigieburn Forest and Ski Basin. The data used in the simulations are midnight-to-midnight precipitation at the Camp Stream gauge, midnight-to-midnight mean daily discharge from Camp Stream, maximum and minimum temperatures at Craigieburn Forest and Ski Basin, and the

TABLE 1: Characteristics of Camp Stream basin.

Elevation Range (m)	1020–1070	1070–1220	1220–1370	1370–1530	1530–1680	1680–1730
Area (km^2)	.06	.25	.26	.26	.09	.02
% Forest	100	90	63	0	0	0
% Tussock	0	10	18	25	20	0
% Scree	0	0	14	75	80	100
% Scrub	0	0	5	0	0	0

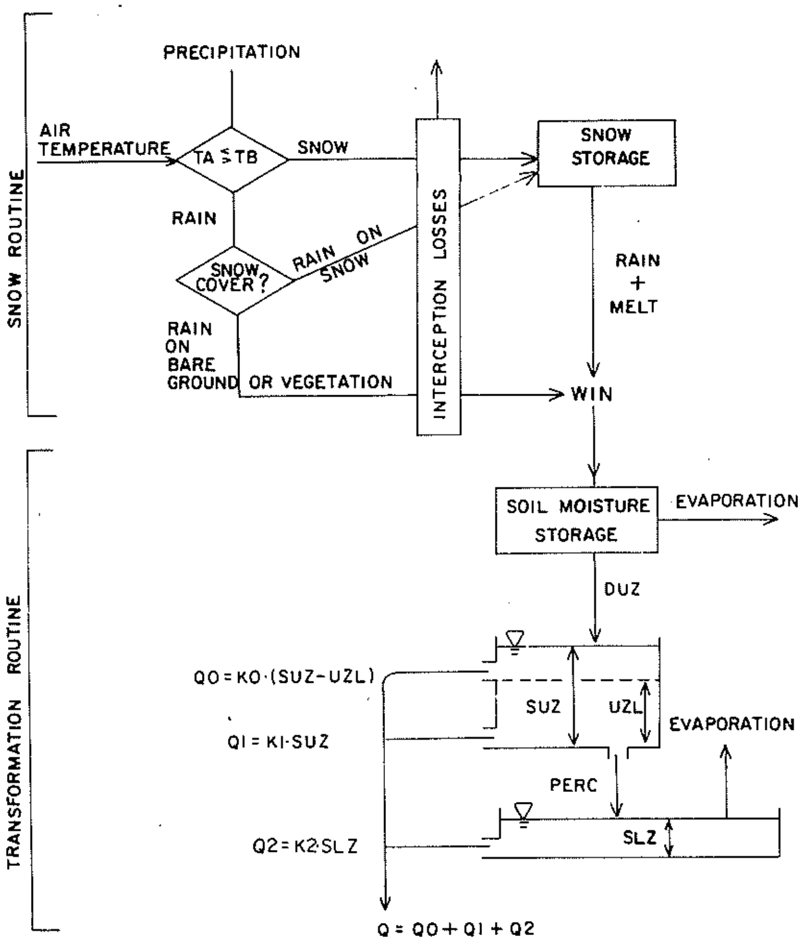

FIG. 2: Schematic of model. Details of the snow and soil moisture routines and interception loss calculations are given in the text.

surface pressure difference measured at noon between Auckland and Invercargill (Fig. 1). Moore and Owens (1984) have shown that this pressure difference is an index, at a site near the Main Divide of the South Island, of wind speed during westerly airflow. Periods of strong westerly winds have been shown by Prowse and Owens (1982) to dominate rapid melt at the Ski Basin site. The mean of the maximum and minimum temperatures at Craigieburn Forest and Ski Basin was linearly interpolated to calculate air temperatures at different elevations.

MODEL DESCRIPTION

The model structure is shown schematically in Figure 2. The disposition of precipitation by the snow routine is according to elevation band and vegetation type, as shown in Table I (scrub is treated as tussock). Precipitaton is treated as rain if air temperature is greater than a threshold temperature, TB, and as snow otherwise. Rain falling onto forest is decreased by thirty percent, the probable interception loss (Rowe; 1975, 1983). Interception losses from snow-free tussock are calculated as twenty percent of gross rainfall (Pearce and McKerchar, 1979). Interception losses of snow are calculated only for forest, and are assumed to be thirty percent of gross snowfall, based upon measurements reported by O'Loughlin (1969).

In non-forested areas, meltwater is retained by the snow pack as free water when snowmelt begins. Runoff can occur once the water retention capacity, WRC (mm free water per mm snow water equivalent), is filled. Refreezing of liquid water is computed as

$$RFR = CRF.(0 - TA) \tag{1}$$

where RFR is the quantity of water refrozen (mm d^{-1}), CRF is a refreezing coefficient (mm $d^{-1} {}^{\circ}C^{-1}$) and TA is air temperature (°C). The influence of snow water processes in the forest was considered to be small enough not to warrant inclusion.

Melt is assumed to occur if air temperature TA is greater than the threshold value TB on days with precipitation, and if air temperature is greater than zero on days without precipitation. Two melt routines were used. One is a simple degree-day factor (DDF) approach in which snowmelt, M (mm d^{-1}), is given by

$$M = MF.TA \tag{2}$$

where MF is a melt factor (mm $d^{-1} {}^{\circ}C^{-1}$). The other is an index energy budget (IEB) model, similar to Anderson's (1973) scheme which calculates melt differently for clear days or days of light rain and for days of heavy rain. Prowse and Owens (1982) and Moore (1983) have shown that sensible heat exchange is the dominant heat input during spring snowmelt at Ski Basin. Net radiation is next most important in fine weather, while the latent heat of condensation exceeds net radiation on rainy days. Accordingly, the energy available for melt, Q_M (W m^{-2}), is calculated from

$$Q_M = \begin{cases} Q^* + Q_H & P < 10 \text{ mm d}^{-1} \\ Q^* + Q_H + Q_E + Q_P & P \geq 10 \text{ mm d}^{-1} \end{cases} \tag{3}$$

where Q^* is net radiation, Q_H and Q_E are the exchanges of sensible and latent heat, Q_P is the sensible heat of precipitation and P is daily precipitation.

Net radiation on clear days is represented by a function which is similar to Anderson's (1973) formula for snowmelt in clear weather:

$$Q^* = RMF.[0.5 + 0.5 \cos (2\pi JD/365)].TA \tag{4}$$

where RMF is a radiation melt factor (W $m^{-2} {}^{\circ}C^{-1}$) and JD is the Julian day number referenced to the summer solstice. On rainy days, shortwave radiation is assumed to be negligible, and the atmosphere is assumed to radiate as a blackbody at the air temperature. Anderson (1973) showed that, under these assumptions, net radiation on rainy days can be expressed in W m^{-2} as

$$Q^* = 4.6.TA \tag{5}$$

if the snow surface is 0°C. Precipitation heat flux is calculated by assuming that the air is saturated, and that the precipitation temperature equals the wet bulb temperature (Anderson 1973); that is,

$$Q_P = \rho.c.P.TA / (86.4 \times 10^6) \tag{6}$$

where ρ is the density of water (kg m^{-3}) and c is the specific heat of water (J kg^{-1} °C^{-1}). The turbulent fluxes of sensible and latent heat are represented by

$$Q_H = C_P.(C1 + C2.DP).TA \tag{7a}$$

$$Q_E = (C1 + C2.DP).(e^* (TA) - 6.11) \tag{7b}$$

where C_P is the psychrometric constant, C1 and C2 are empirical constants, DP is the Auckland-Invercargill pressure difference (mb) and e* (TA) is the saturation vapour pressure at air temperature (mb). DP is included as an index of wind speed to improve the model's performance for strong regional advective conditions. DP was set to zero when it was negative, although a negative pressure difference usually coincided with non-melt situations. The daily total of energy used in snowmelt is converted from W m^{-2} to mm d^{-1} by multiplying by .259.

Rates of snowmelt under forest cover have not been investigated in New Zealand. Therefore, snowmelt rates for forested areas are calculated from

$$M_F = 1.5.TA + P_F.TA/79.7 \tag{8}$$

where M_F is in mm d^{-1}, P_F is the net precipitation in the forest in mm d^{-1} and the melt factor of 1.5 is an intermediate value of melt factors for forested areas as given by U.S. Army Corps of Engineers (1956) and Bengtsson (1982). The second term on the right hand side represents the melt caused by the sensible heat of precipitation, which can be substantial (Fitzharris *et al.*, 1980).

Net rainfall and snowmelt calculated by the snow routine become input to the lumped transformation routine. This water input, WIN, first enters the soil moisture routine where a portion of it, DUZ (mm), passes through to the upper subsurface reservoir. DUZ is calculated from

$$DUZ = \begin{cases} WIN.(SM/FC)^{BETA} & SM < FC \\ WIN & SM \geq FC \end{cases} \tag{9}$$

where SM is soil moisture storage (mm), FC is field capacity (mm) and BETA is a dimensionless parameter. Water input is added to the soil moisture routine and partitioned mm by mm, as this is more realistic than allowing the antecedent soil moisture to govern the disposition of the entire day's input.

Evaporative losses from the soil moisture zone are calculated from

$$E = \begin{cases} PET.(SM/CL).VFRAC & SM < CL \\ PET.VFRAC & SM \geq CL \end{cases} \tag{10}$$

where E is evaporative loss (mm), CL is the soil moisture content below which evaporation is impeded, VFRAC is the fraction of the catchment from which evaporation is occurring and PET is potential evapotranspiration (mm). Monthly evapotranspiration rates were derived by estimating the annual total and then distributing this total through the year in proportion to the monthly distribution of pan evaporation at Lake Coleridge, which is 10 km south of the Craigieburn Range and has an elevation of 533 m (Bowden, 1983). Annual potential evapotranspiration in Camp Stream catchment is estimated at 400 mm, based upon a discontinuous record of pan estimates at Craigieburn Forest given in unpublished New Zealand Forest Service reports, and a water-balance estimate of a small (22 ha) catchment adjacent to the Craigieburn Forest site. The area from which evaporation occurs is calculated by assuming that evaporation occurs from the forest all year, but from non-forested areas only when free of snow.

The portion of the daily water input which passes through the soil moisture zone, DUZ, is added to the upper-zone storage, SUZ (mm). If SUZ exceeds a threshold value UZL (mm), a runoff component Q0 is calculated as

$$Q0 = K0.(SUZ - UZL) \qquad (11)$$

where K0 is a recession constant (d^{-1}). Q0 is then subtracted from SUZ and a second runoff component, Q1, is calculated from

$$Q1 = K1.SUZ \qquad (12)$$

where K1 is another recession coefficient. After Q1 is subtracted from SUZ a further quantity of water, PERC, is lost from the upper zone to recharge the lower subsurface storage, SLZ. If SUZ is less than PERC, all of SUZ is lost to the lower reservoir.

After the recharge from the upper zone is added to SLZ, evaporation from wet areas is calculated as PW.PET, where PW is the fraction of the catchment covered by lakes and wet areas, and subtracted from SLZ. The release of water from the lower zone is analogous to the groundwater component of streamflow, and is

$$Q2 = K.SLZ \qquad (13)$$

where K2 is a recession constant.

The runoff components Q0, Q1 and Q2 are summed for each day and converted from mm d^{-1} to $l\ s^{-1}$. Channel routing, as is incorporated into the HBV model, was not carried out because the time of concentration in Camp Stream is much less than the daily time step used in the simulations.

MODEL CALIBRATION

Periods in which any of the required input data had large gaps or were of doubtful quality were discarded, leaving only 1981 and up to the end of November 1982. The year 1981 was arbitrarily selected as the calibration period. Calibration was performed by visual inspection of observed and modelled hydrographs and by comparison of several numerical indices, including the model efficiency of Nash and Sutcliffe (1970) (defined in Table 2), the cumulative error, and regression relationships between modelled and observed streamflow.

The snow routine has four or six parameters, depending upon whether the degree-day factor or the index energy budget melt routine is used, while the transformation routine has nine. In addition, initial values for the three state variables in the transformation routine (SM, SUZ and SLZ) must be specified.

Initial values for the snow routine parameters were found by calibrating the routines against snow course data from Alan's Basin. Camp Stream basin is well-drained and has no lakes, so PW was set to zero. Values for K0, K1, K2 and PERC were found through hydrograph analysis, as described by Bergstrom and Forsman (1973) and Bergstrom and Jonsson (1976). Initial estimates of BETA, FC, CL and UZL were made in consideration of values reported by Bergstrom and Forsman (1973), Bergstrom (1975), Bergstrom and Jonsson (1976) and Svensson (1977). Initial values for SUZ and SLZ were set to correspond to the observed discharge on the first day of the simulation period, and soil moisture was set to field capacity. Applying a snowfall correction factor and varying precipitation with elevation did not improve model fit.

During calibration, a problem was noted in fitting the recession limbs. Adjusting parameters to maximize the model efficiency, E, tended to fit the larger peaks at the expense of fitting the recessions (Tables 2 and 3; Fig. 3). When parameters were adjusted to fit the recessions, E decreased markedly, and the model tended to respond too rapidly to smaller runoff events (Fig. 3). This problem is related to the significant cumulative error: the modelled runoff exceeded observed runoff during the calibration period by approximately 300 mm (Table 2). This cumulative error is equivalent to fifteen percent of the gross precipitation, or thirty percent of observed runoff, and is almost the same magnitude as the estimated evapotranspiration. The choice of parameters in effect dictates when the excess water is routed through the basin.

MODEL VALIDATION

Performance indices for the 1982 simulations using the index energy budget and degree-day factor models are shown in table 2. The degree-day factor model provided the better fit according to the indices, but neither performed satisfactorily. Both simulations again show a positive cumulative error. Figure

TABLE 2: Performance indices for runoff simulations.

Run	1	2	3	4	5
CE	300	297	308	223	218
RMSE	20.0	20.8	27.0	25.3	24.4
E	0.76	0.74	0.56	0.32	0.36
R^2	0.81	0.79	0.72	0.52	0.53
A	14.0	1.2	9.5	13.2	13.8
B	0.84	0.83	1.0	0.79	0.77

CE = cumulative runoff error (mm)
RMSE = root mean square error (l/s)
E = model efficiency = 1 — $(RMSE/SD)^2$ (SD = standard deviation of observed flows)
R^2, A, B = coefficient of determination, intercept and slope of regression of modelled against observed flow

Run
1 Calibration run, degree-day factor model, 1981
2 Calibration run, index energy budget model, 1981
3 Calibration run, index energy budget model, 1981
4 Validation run, index energy budget model, 1982
5 Validation run, degree-day factor model, 1982

TABLE 3: Parameter values for runoff simulations.

Run	1	2	3	4	5
TB	1.2	1.2	1.2	1.2	1.2
C1	—	10.0	10.0	10.0	—
C2	—	1.6	1.6	1.6	—
RMF	—	0.0	0.0	0.0	—
MF	4.0	—	—	—	4.0
WRC	0.0	0.0	0.0	0.0	0.0
PW	0.0	0.0	0.0	0.0	0.0
CRF	0.0	0.0	0.0	0.0	0.0
FC	150	150	150	150	150
CL	90	90	90	90	90
BETA	8.0	8.0	8.0	8.0	8.0
K0	0.5	0.5	0.5	0.5	0.5
K1	0.06	0.06	0.10	0.06	0.06
K2	0.025	0.025	0.025	0.025	0.025
PERC	3.0	3.0	3.0	3.0	3.0
UZL	75.0	75.0	50.0	75.0	75.0

Note: run numbers correspond to those in Table 2.

4 illustrates the simulations; visually, there is little to choose between the two. Neither simulation reproduces the two hydrograph peaks during the snowmelt period between 28 October and 9 November; conditions leading up to and during this period are discussed by Moore (1983) and Moore and Owens (1984). Figure 5 shows the snow storage in the open portions of the upper five elevation zones as modelled using the index energy budget melt routine. The accumulation at 1105 m, 1605 m and 1705 m appears reasonable when compared with Craigieburn Forest and Ski Basin climate records and personal observations, but the snow storage at 1295 m and particularly at 1450 m melts prematurely. Consequently, the modelled area contributing snowmelt runoff during the 28 October to 9 November melt period is underestimated, producing an underestimate of runoff.

An alternative approach has been taken to modelling the snowmelt period. Estimated snow-lines based on field observations during the melt period were used to calculate the snow-covered area for each day. To simplify the calculations, snowmelt under the beech forest has been neglected; the area contributing snowmelt was therefore derived from a plot of the non-forested area in Camp Stream above a given elevation against the elevation. The mean daily temperature at the median elevation of the snow-covered area was used in the snowmelt calculations. Net precipitation was calculated by subtracting interception losses from rain falling onto the forested portion of the catchment. The daily totals of snowmelt averaged over the basin plus net precipitation were input to the transformation routine. As before, soil moisture was initially set to field capacity, and hydrograph analysis indicated that the upper storage was almost empty. Therefore, SUZ was set to 0 mm and SLZ to 101 mm, to correspond to the discharge of 27.5 $l\ s^{-1}$ on 28 October. When the model

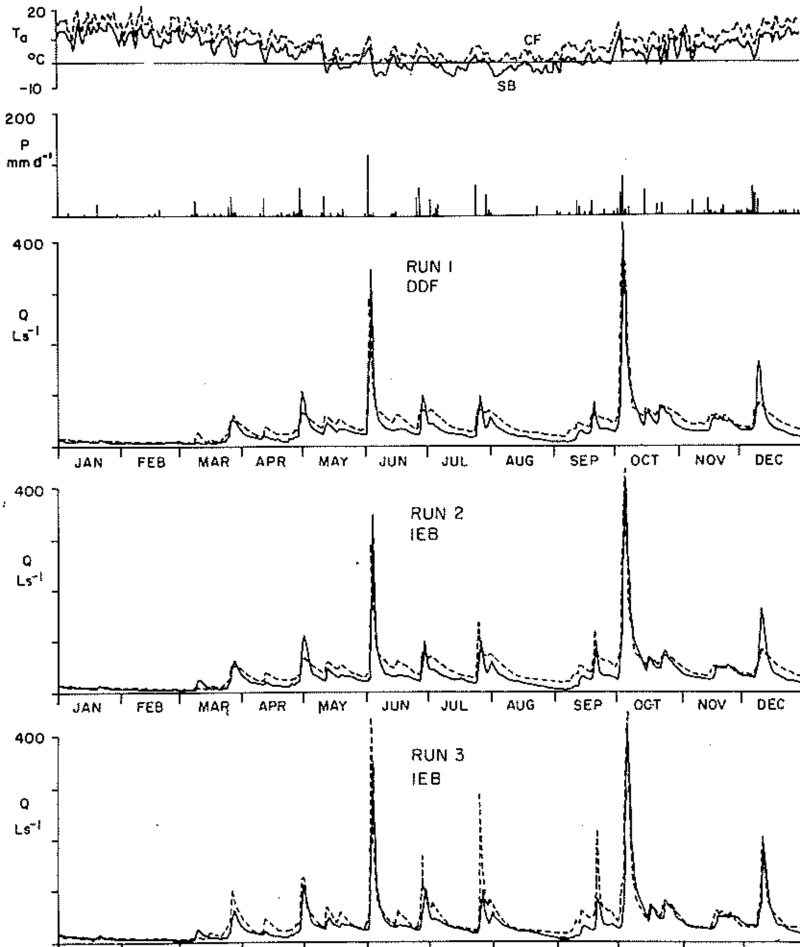

FIG. 3: Calibration simulations, 1981: solid line — observed; dashed line — modelled. (Run numbers correspond to those in Table 2).

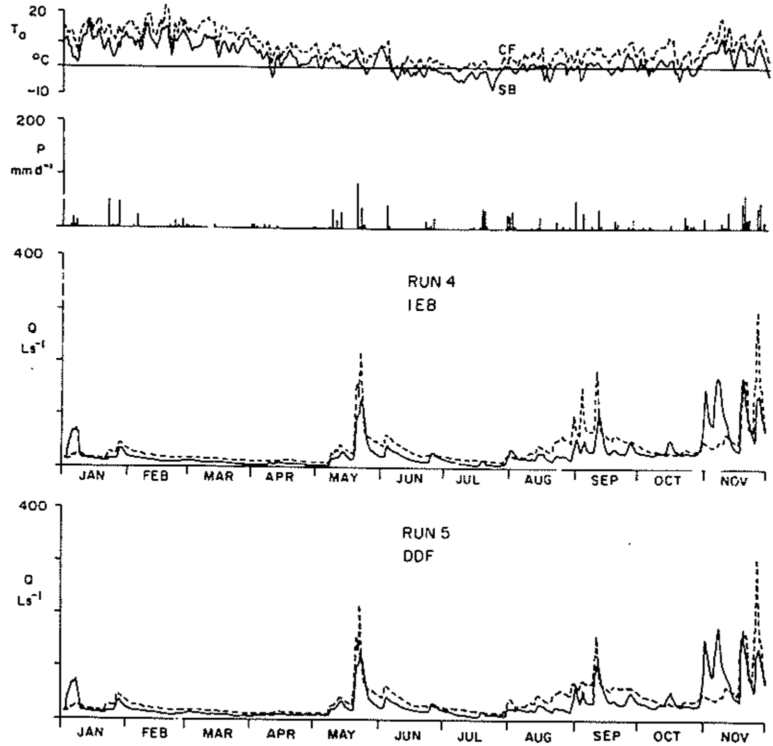

FIG. 4: Validation simulations, 1982.

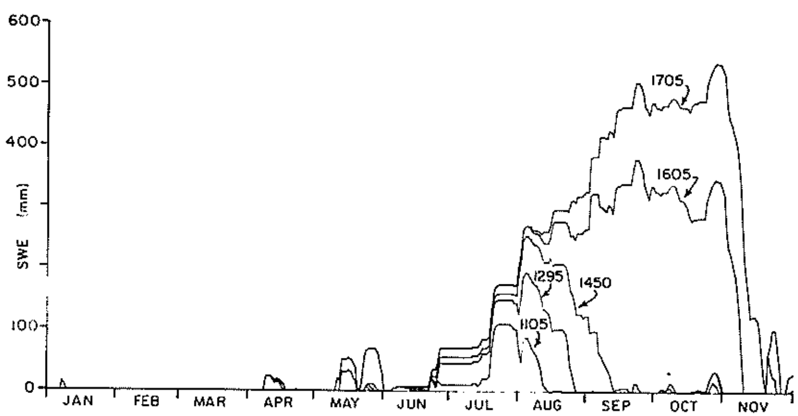

FIG. 5: Simulated snow accumulation in four elevation zones, 1982. (SWE is snow water equivalent).

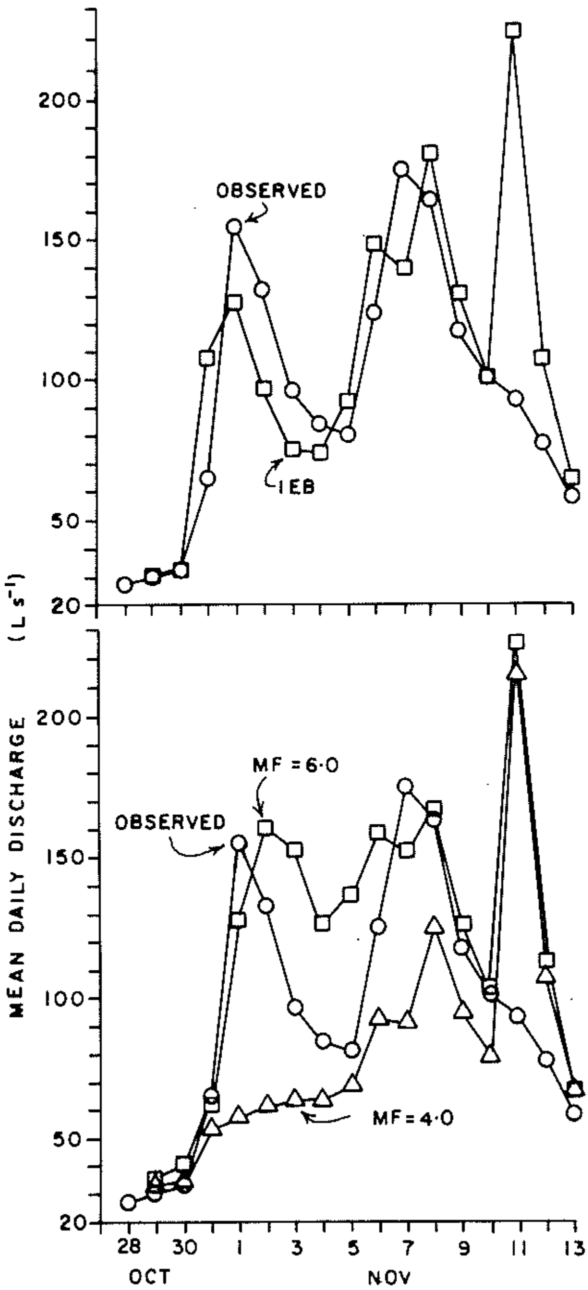

FIG. 6: Modelled and observed discharges, spring 1982.

was run, however, basin response to the first rain-on-snow pulse was far too damped, suggesting either that the upper zone was not initially empty, or that either or both of UZL and K1 were inappropriate. Therefore, SUZ and SLZ were set arbitrarily such that each reservoir contributed equally to the discharge on 28 October.

Figure 6 shows the results of simulations using the index energy budget melt routine and the degree-day factor model, with two values for the melt factor, and Table 4 shows the snowmelt calculations. The index energy budget model provides a better reproduction of the variability of runoff. The degree-day factor model does not adequately account for short-term melt variations during changing weather situations. All simulations display an unrealistic response to the 32.5 mm of net rainfall on 11 November. This over-response is not as evident in the simulations shown in Figure 4.

DISCUSSION

Several problems were encountered in this modelling exercise. One was the substantial volume error, which made calibration of the response function parameters difficult. The volume error could be caused by one or more factors: underestimation of evaporation and interception losses, unrepresentative precipitation data, erroneous streamflow data, unaccounted loss from blowing snow and subsurface losses. The precipitation gauge probably has a negative bias because it has no wind shield and is only 60 m above the median basin elevation, so overestimating precipitation appears unlikely to account for the volume error. The importance of the other possible sources of error is difficult to assess without further information.

Another problem was the misclassification of snow events below 1500 m. Analyses by O'Loughlin (1969) and Prowse (1981) indicate that the magnitude of temperature variations preceding, during and following snow storms in the Craigieburn Range decreases with elevation, with freezing-level fluctuations being confined mainly to the zone below 1500 m. Hence, snow events are more likely to be misclassified by using daily mean temperature in this zone than at higher elevations. Large portions of Camp Stream and other South Island catchments lie below 1500 m, so precipitation and temperature data with a finer temporal resolution are required for successful modelling of this environment.

The index energy budget model reproduced the variation of melt during the spring 1982 melt period more faithfully than the degree-day factor model. This finding supports the inference from energy balance considerations that a wind-speed index is required in melt calculations for the South Island. In this study, the Auckland-Invercargill pressure difference was used as the wind-speed index because it is readily available, and the time frame of wind speed measurements as Ski Basin does not coincide with those of the precipitation and streamflow data. However, the pressure difference may not be the most appropriate index for use in the Craigieburn Range, which lies in the lee of the Main Divide mountains.

Some problems were encountered in transforming the calculated melt and net precipitation into streamflow. The lumped representations of soil moisture and upper-zone storage may be inaccurate. During the snowmelt period, most

TABLE 4: Snowmelt calculations for spring 1982 melt period.

Date	SL	SCA	T	DP	P	M1	M2	M3
Oct 28	1200	.47	1.9	6.2	0.0			
29	1200	.47	5.2	6.1	0.0	20.8	31.2	14.9
30	1200	.47	4.2	8.8	0.0	16.8	25.2	14.7
31	1230	.46	6.3	20.1	23.1	25.2	37.8	79.9
Nov 1	1260	.44	7.4	13.0	0.0	29.6	44.4	33.1
2	1300	.42	7.1	5.6	0.0	28.4	42.6	19.5
3	1350	.38	6.4	3.1	0.0	25.6	38.4	13.9
4	1400	.32	5.6	7.4	0.0	22.4	33.6	17.7
5	1425	.28	8.9	16.8	0.0	35.6	53.4	47.6
6	1450	.24	12.1	22.7	0.0	48.4	72.6	81.3
7	1450	.24	10.4	16.5	0.0	41.6	62.4	54.9
8	1500	.15	6.1	15.0	14.5	24.4	36.6	62.2
9	1525	.10	9.1	17.6	5.8	36.4	54.6	50.4
10	1525	.10	7.5	19.1	5.8	30.0	45.0	44.1
11	1525	.10	1.6	6.6	37.6	6.4	9.6	10.0
12	1525	.10	1.3	8.6	2.9	5.2	7.8	4.5
13	1525	.10	3.5	0.9	0.0	14.0	21.0	5.8

SL — snow-line elevation (m)
SCA — snow covered area (km^2)
T — temperature at median elevation of snow zone (°C)
DP — Auckland-Invercargill pressure difference (mb)
M1 — melt (mm/d) calculated with MF = 4 mm/ (d°C)
M2 — melt (mm/d) calculated with MF = 6 mm/ (d°C)
M3 — melt calculated with index energy budget model

runoff was generated on the upper portions of the basin, while the soil dried out in areas below snow-line. When rain fell on 11 November, the response of the snow-free areas would have been damped as the soil moisture deficit was satisfied. However, the lumped transformation model appears to have misrepresented the equivalent antecedent moisture status of the basin, producing too rapid a response. Ferguson (1984) found that rain inputs were routed less correctly than snowmelt in his application of a simple lumped model in the Cairngorm mountains. He postulated that failure to account for soil moisture conditions on areas below snow-line caused the overestimation of basin response.

Inhomogeneity of response could also explain the problems with setting initial conditions for simulating the 1982 snowmelt period. The recession constants were derived from a year dominated by rainfall, which inputs water over the entire catchment. However, snowmelt occurred mainly in the upper portion of the catchment dominated by tussock and scree, where runoff is generated differently compared to the lower, forested portions of the catchment.

The year used for calibration, 1981, was a light snow year, and adequate activation of the snow routine parameters may not have occurred. Bergstrom (1975) found that four years of data were insufficient to fit stable parameter values in a Swedish catchment. Sorooshian (1983) stated that calibration data should be chosen to maximise hydrological variability; therefore, at least one light and one heavy snow year should be used for calibration.

CONCLUSIONS

A runoff model incorporating a distributed snow routine and a lumped transformation routine was developed for conditions in non-glacierized mountainous basins in the eastern South Island of New Zealand, and was applied in the Camp Stream catchment. A major problem was the over-estimation of runoff. The magnitudes of the water-balance components in the South Island mountain regions, particularly evaporation and subsurface basin outflow, should be investigated to improve the model. Mean daily air temperature is inadequate for discriminating rain from snow at elevations lower than about 1500 m. Precipitation and temperature data with a resolution of three hours or less may be required to simulate snow accumulation. Further modelling work on this environment should include distributed transformation routines and the refinement of snowmelt indices. In particular, the use of alternative wind-speed indices, such as wind speeds from upper air-pressure, should be explored.

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