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MODELLING ALPINE SNOW ACCUMULATION AND ABLATION USING DAILY CLIMATE OBSERVATIONS

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ABSTRACT

A simple snow accumulation and ablation model using daily observations of temperature and precipitation was used to simulate snow storage at a snow course. Low-altitude precipitation measurements appear to provide a good index of high-altitude precipitation when averaged over the accumulation season. Snow storage was reasonably reproduced when temperatures measured at a site above the seasonal snow-line were used. However, the simulation quality was less satisfactory when low-altitude temperatures were used, because deviations of the actual from the assumed lapse rate make it difficult to distinguish rain from snow. If no high-altitude temperature measurements are available, storm snow-line elevations may be used to determine the phase of precipitation. Application of these approaches may at least partially overcome lack of data on snow storage in New Zealand mountains.

INTRODUCTION

Many conceptual runoff models incorporate snow accumulation and ablation algorithms which use meteorological input to simulate the snow storage within a basin. The parameters in these algorithms are usually optimized by obtaining the best fit to an observed hydrograph. However, this comparison is indirect, so several workers have compared modelled water equivalents of snow to water equivalents measured during the accumulation season (for example Anderson, 1976; Speers *et al.*, 1979; Huber, 1983). If the model performs adequately, it can then be used to increase the temporal resolution and extend the record of measurements (Woo, 1972). This would be an advantage in New Zealand where only a small number of snow courses have been measured in relation to the winter snow-covered area (Prowse and Owens, 1982), and there is considerable cost and logistical difficulty in undertaking such measurements (Gillies, 1964). Another advantage is that parameter values in a runoff model's snow routine can be assigned reasonable initial values independently of calibration with streamflow data. In this paper, modelled snow storage is compared to data from a snow course in the Craigieburn Range, and the utility and limitations of this approach are assessed for New Zealand conditions and data availability.

DATA BASE

The snow course data used in this study were collected by the Forest Research Institute in Alan's Basin at an elevation of 1750 m in the Craigieburn Range

(Fig. 1), which lies 20 km east of the Main Divide of the South Island. Drifting of snow is uncommon on the snow course, which lies on a smooth lee slope of approximately 28° (Morris and O'Loughlin, 1965). Snow depth and density were sampled at points 20 m apart on a transect which followed the local contour. Density was measured with an Italian CN-1 sampler for the first years, but from 1968 onwards a Mount Rose sampler was used (C. L. O'Loughlin, pers. comm.).

Snow course data are subject to both measurement and sampling errors, each of which can have random and systematic components. The densities used for calculating water equivalent were not corrected for sampler bias by the Forest Research Institute; therefore, the water equivalents have been reduced by eight percent for use in this study, following the results of Work *et al.* (1965) and Jordan (1978). The variabilities of both snow depth and density are probably much larger than the measurement imprecision, so errors in water equivalent should be caused mainly by sampling errors. Estimates of the sampling errors indicate that the mean values for snow-course water equivalents are precise to within ±70 mm at the ninety-five percent confidence interval (Moore, 1984).

The climate data were collected at the Ski Basin and Craigieburn Forest stations (Fig. 1), which have elevations of 1550 m and 914 m, respectively. Maximum and minimum daily temperatures from both stations were used. However, only daily precipitation totals from Craigieburn Forest were used because the effects of wind and snow accumulation around the gauge render the precipitation record at Ski Basin unreliable.

Maximum and minimum temperatures and precipitation are read at 0900 NZST at Craigieburn Forest and between 1000 and 1100 at Ski Basin. The minimum temperature is assumed to have occurred on the morning of the day of observation and is entered in the climate record for that day. The maximum temperature and precipitation, on the other hand, are entered in the record for the previous day. Therefore precipitation and maximum temperature for one day and the minimum temperature for the following day were used to calculate synchronous daily precipitation and mean daily temperature. Temperatures from both Craigieburn Forest and Ski Basin were extrapolated to the snow course elevation using a lapse rate of 6.5°C km^{-1}, which is a commonly-assumed lapse rate in mountainous areas (Obled and Harder, 1979; Young, 1982; Bagchi, 1983) and is similar to the mean storm lapse rate in the Craigieburn Range, which Prowse (1981) calculated to be 6.2°C km^{-1}. It is also close to the average value of 6.0°C km^{-1} reported by Coulter (1967) for Black Birch, Marlborough.

MODEL DESCRIPTION

Precipitation and mean temperature for each day are read in and extrapolated to the snow course elevation. The temperature at the snowcourse is assessed as described above, and the precipitation by multiplying by a constant correction factor which accounts for the change of new snow accumulation with elevation and includes effects such as the increase of precipitation and decrease of evaporation with increases of elevation. In addition, as most precipitation in this area occurs in north-west storms (Prowse, 1981), the amount of wind-redistributed snow would be expected to increase towards the ridge line on the east side of the range. If the extrapolated air temperature is less than or equal to a threshold temperature, the precipitation is considered to be snow and is added to the snow pack; otherwise, it is treated as rain.

Melt is assumed to occur on days with no precipitation and air temperature greater than 0° C, or on days with precipitation and air temperature greater

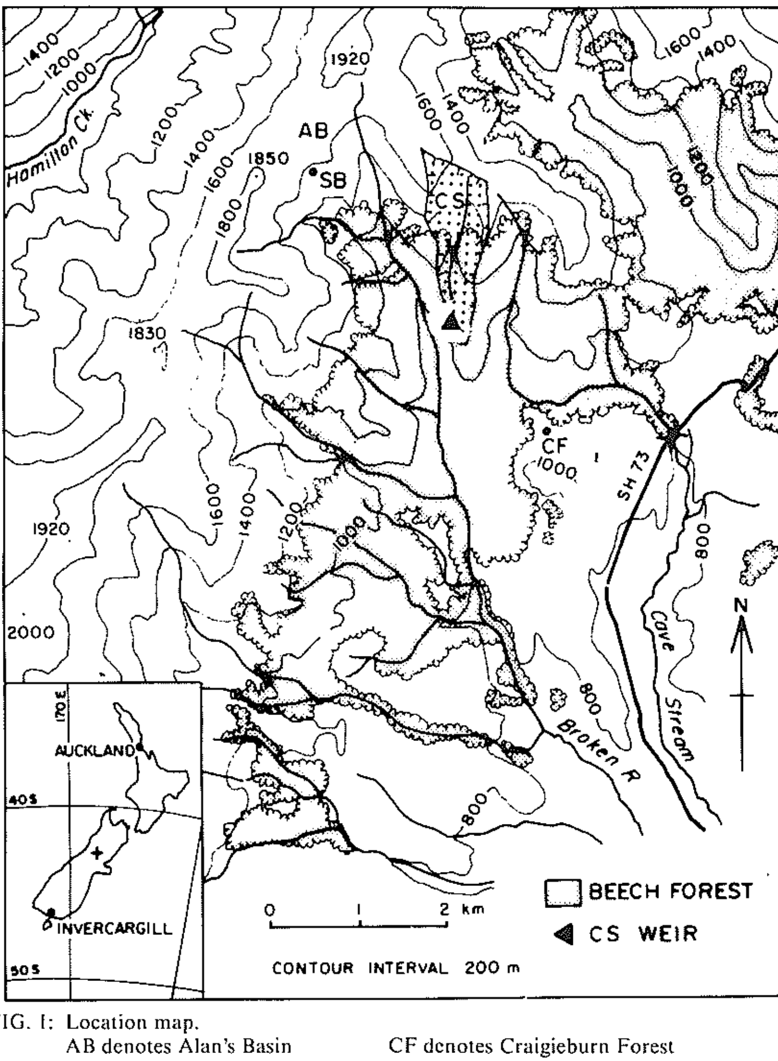

FIG. 1: Location map.
AB denotes Alan's Basin
SB denotes Ski Basin
CF denotes Craigieburn Forest
CS denotes Camp Stream

than a threshold temperature. Snowmelt is calculated from a simple degree-day method, which is

$$M = MF.TA \tag{1}$$

where M is snowmelt in mm d^{-1}, MF is the degree-day factor (mm $d^{-1}{}^{\circ}C^{-1}$), and TA is the extrapolated air temperature.

Snowmelt and rainfall are retained within the snow pack until a water-retention capacity (in mm water per mm snow water equivalent) is exceeded. Once the retention capacity is filled, snowmelt and rain falling onto the snow pack are considered to run off. If the temperature drops below 0°C, freezing of the liquid water content is computed as

$$RFR = CRF .(0 - TA) \tag{2}$$

where RFR is the quantity of water (mm d^{-1}) which is refrozen and added to the snow pack and CRF is a refreezing coefficient (mm $d^{-1}{}^{\circ}C^{-1}$).

The model thus has five parameters which must be determined: precipitation correction factor, threshold temperature, degree-day factor, refreezing coefficient and water-retention capacity. The method for determining thse is described in the next section.

PROCEDURE

The models were calibrated by mapping the response surface of an objective goodness-of-fit criterion. The advantage of this procedure over automatic optimization methods is that local optima are more easily recognized and avoided. Fixed intervals of 0.1 for threshold temperature, refreezing coefficient, water-retention capacity and precipitation correction factor and 0.5 for degree-day factor were used as a compromise between maximizing the extent of both the parameter space examined and the resolution, while minimizing computational time. The parameter values found may not be optimal being constrained by the intervals used. The criterion employed is Nash and Sutcliffe's (1970) model efficiency, given by

$$E = 1 - (RMSE/SD)^2 \tag{3}$$

where RMSE is the root mean square error between modelled and observed series and SD is the standard deviation of the observed series. Possible values of E range from unity, indicating a perfect fit, to negative infinity for poor fits. A value of E equal to zero implies that the mean of the observed series is as efficient a predictor as the modelled values. This criterion is equivalent to minimizing the squared error between observed and modelled values. Model fitting on the basis of least-squares criteria can, however, result in biased parameters because of the statistical properties of input data and model output (Sorooshian, 1983). The procedure is used here to provide an objective comparison of model performance during different simulations.

To investigate the sensitivity of the simulation to the sampling period used, the model was first calibrated using the 1967–68 data and then run on the 1969–73 data using the 1967–68 optimal parameters. The model was then calibrated using the entire data sequence to investigate the fit of the model to the snow course data. Each simulation was carried out using both Craigieburn Forest and Ski Basin temperatures to determine errors involved in extrapolating data from low altitudes to simulate alpine snow storage.

TABLE 1: Optimal parameters from snow accumulation simulations.

	1967-68	1967-73
Craigieburn Forest Temperatures		
MF	4.0	4.0
TB	2.1	2.2
PCF	1.2	1.1
Ski Basin Temperatures		
MF	8.0	6.0
TB	1.2	1.2
PCF	1.3	1.3

MF — degree-day factor (mm $d^{-1} {}^{\circ}C^{-1}$)
TB — threshold temperature (°C)
PCF — precipitation correction factor
Note: Water-retention capacity and refreezing coefficient set to zero.

RESULTS

Table 1 sumarizes the optimal parameters found by the mapping procedure. The parameters water-retention capacity and refreezing coefficient were included for the calibration with Ski Basin temperatures. However, model performance decreased whenever they were not set to zero, so they were set to zero for all other runs. The parameters governing snow accumulation, threshold temperature and precipitation correction factor are stable with respect to calibration period when Ski Basin temperatures were used. In addition, the precipitation correction factor is physically realistic in that mean annual precipitation at Ski Basin is approximately 1.24 times greater than at Craigieburn Forest (McCracken, 1980). These parameters are less stable and less realistic when Craigieburn Forest temperatures were used. The values 4.0 to 8.0 mm $d^{-1} {}^{\circ}C^{-1}$ for degree-day factor are higher than those reported elsewhere (Male and Gray, 1981), which may reflect the importance of sensible-heat transfer in snowmelt in this environment (Prowse and Owens, 1982).

Figure 2 shows portions of the model efficiency topography surrounding the optimal values for the 1967-68 calibration simulations. The same intervals were used for mapping each variable so the model sensitivity can be compared for the simulations using Craigieburn Forest and Ski Basin temperatures. Interaction between each pair of variables is apparent when either Craigieburn Forest or Ski Basin temperatures were used, but the sensitivity is greater when Craigieburn Forest data were used. This interaction implies that an incorrect setting of one parameter can, to some extent, be compensated for

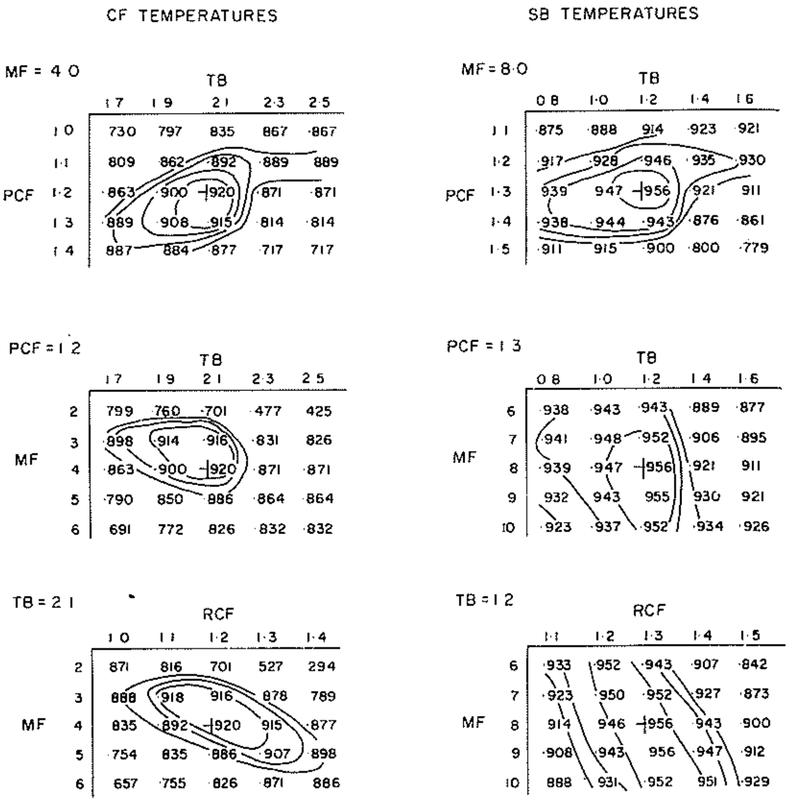

FIG. 2: Portions of model efficiency topography for 1967–68 calibration. Isopleth interval is 0.1; abbreviations explained on Table 2.

during calibration by distortion of other parameter values. However, the optima in both cases are reasonably well-defined.

Table 2 summarizes the goodness-of-fit indices for the calibration and verification periods. The simulation using Craigieburn Forest temperatures fitted poorly, and showed the greatest deterioration in fit between the 1967–68 and 1969–73 periods. The root mean square errors for the simulations using Ski Basin temperatures are approximately 50 to 70 mm, roughly the same as the sampling errors of the snow course data.

The modelled and observed water equivalents are illustrated in Figures 3 and 4. The simulations using Craigieburn Forest data are visually poorer fits than the other simulations, particularly during 1970. These plots show that the quality of simulation varies between years for all simulations. The variation is due to differences in weather patterns, which produce departures from the constant values of precipitation correction factor, threshold temperature and degree-day factor.

TABLE 2: Performance indices from snow accumulation simulations.

	1967–68	1969–73	1967–73
Craigieburn Forest Temperatures			
E	.92	.69	.82
R^2	.93	.75	.83
A	33	75	54
B	.86	.86	.83
CE	318	1423	1301
RMSE	65	105	90
Ski Basin Temperatures			
E	.96	.87	.94
R^2	.96	.90	.94
A	–3.6	1.3	4.0
B	.98	.85	1.0
CE	–180	–1094	299
RMSE	48	68	53

E — model efficiency (defined in text)
CE — cumulative error (mm) = $(Y_i - X_i)^2$
RMSE — root mean square error (mm)
R^2 — coefficient of determination of linear regression of modelled on observed values
A, B — intercept and slope of regression

DISCUSSION

Low-altitude precipitation measurements appear to provide a good index of high-altitude precipitation when averaged over the accumulation season. However, model performance is substantially better when high-altitude temperatures rather than when low-altitude temperatures were used. The poor performance when using low-altitude temperatures is due to marked and systematic deviations of the lapse rates of near-surface air temperature from the assumed rate of 6.5°C km^{-1} and the resultant errors in extrapolating temperature from a low-altitude station.

Lapse rates have been calculated from the daily observations at Craigieburn Forest and Ski Basin for the period 1972 to 1981, inclusive. Two-way analyses of variance using a fixed-effects model indicate that precipitation class, month,

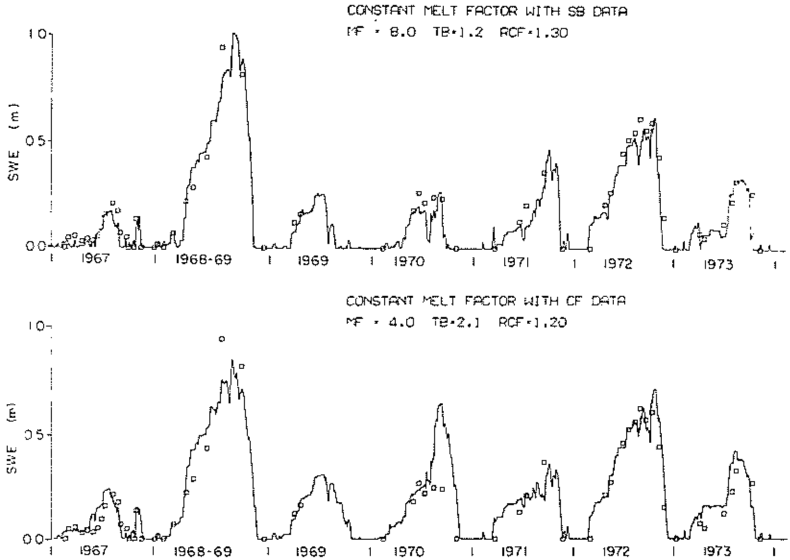

FIG. 3 Observed (squares) and modelled (line) snow accumulation, 1967–68 optimal parameters. Simulation period is May to January for 1968–69, May to December for other years. SWE — snow water equivalent.

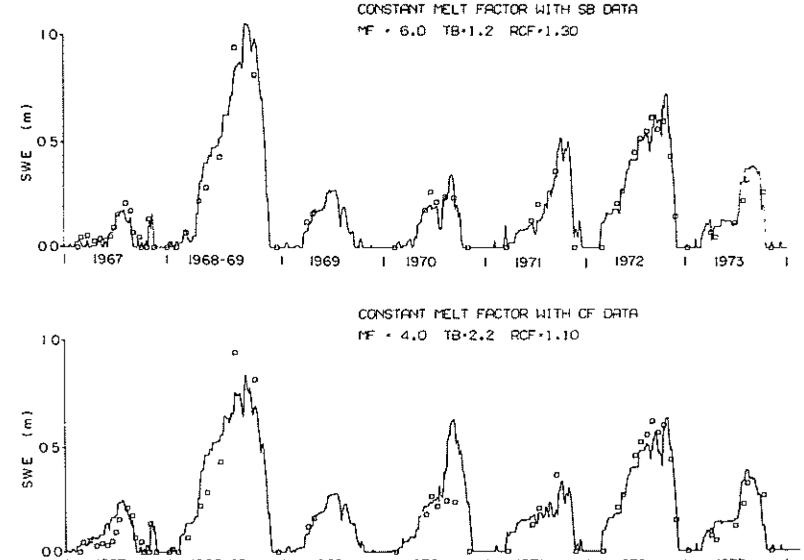

FIG. 4: Observed (squares) and modelled (line) snow accumulation, using 1967-73 optimal parameters. Simulation period is May to January for 1968–69. May to December for other years. SWE — snow water equivalent.

and the interaction effect are all significant at greater than the ninety-five percent level for the lapse rates of maximum, minimum and mean temperatures. For example, Figure 5 shows the calculated lapse rates of mean temperature, by month and precipitation class. Although not shown in Figure 5 for reasons of clarity, variations in the lapse rates are substantial, even when stratified by precipitation class and month: the standard deviations in a given category range from 1.5 to 4.0°C km^{-1}.

These results illustrate the statement of Charbonneau *et al.* (1981) that errors in extrapolating data, particularly temperature, can have a critical effect on simulations. The transition temperature and precipitation correction factor are unstable when using low-altitude temperatures, making it difficult to discriminate rain from snow. If no high-altitude temperature data are available, snow storage simulations may be improved by using new snow-line elevations following storms to determine the phase of precipitation at a given elevation, rather than an extrapolated temperature.

The inclusion of parameters which govern water retention and refreezing did not improve model fit. One reason may be that snow-water phenomena were unimportant given the temporal resolution of the data, and the inherent

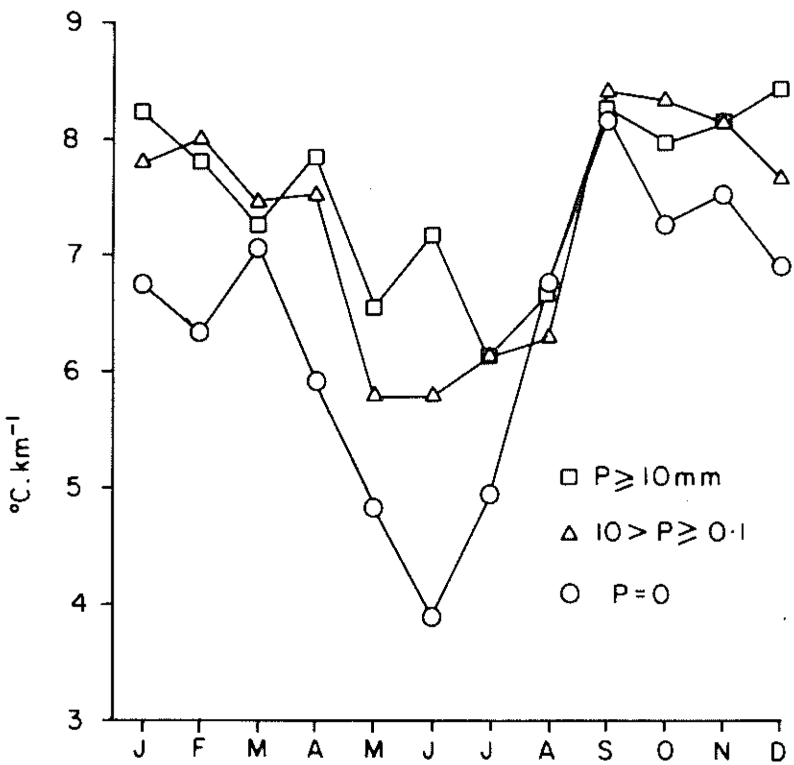

FIG. 5: Lapse rates of mean daily temperature between Craigieburn Forest and Ski Basin for three precipitation classes, 1972 to 1981.

errors in the data and in the simplified representations used in the model. However, if this snow accumulation and ablation model was to be included in a runoff model, non-zero values of refreezing coefficient and water-retention capacity may be required to optimize the fit between modelled and observed runoff.

The simulation results suggest that, where high-altitude meteorological information is available, it may be possible to estimate amounts of snow accumulation. However, despite recent increases in winter observations at skifields, the number of climate stations above 1000m is still quite small, and more high-elevation observations of precipitation and temperature are needed. In addition, if the model is to be applied to markedly different climatic regions or to considerably higher altitudes, it should be calibrated by snow-course measurements.

CONCLUSION

Low-altitude precipitation measurements appear to provide a good index of high-altitude precipitation when averaged over the accumulation season. Snow storage was reasonably reproduced when temperatures measured at a site above the seasonal snow-line were used. However, the simulation quality was less satisfactory when low-altitude temperatures were used, because of marked and systematic deviations of the actual from the assumed lapse rate, and the consequent problem of distinguishing rain from snow. If no high-altitude temperature measurements are available, the use of observations of storm snow-line elevations to determine the phase of precipitation may improve simulation quality. Application of these approaches should help to overcome lack of data on snow accumulation in mountainous areas of New Zealand.

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