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A VARIABLE LAPSE RATE SNOWLINE MODEL FOR THE REMARKABLES, CENTRAL OTAGO, NEW ZEALAND

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ABSTRACT

This paper investigates the computer simulation of winter snowlines on mid-latitude maritime mountain ranges. In previous studies the relationships between snow accumulation and ablation and the more commonly recorded climatic factors, temperature and precipitation, have been investigated. In this paper the relevance of these relationships to the problem of obtaining estimates of winter snowlines is examined. Temperature extrapolation using fixed linear lapse rates is identified as a major source of error in previous studies. Simulations based on measured and interpolated temperature data from three high altitude meteorological stations on the Remarkables, Central Otago, show that the standard modelling routines used are essentially sound, allowing daily snowlines to be modelled with reasonable accuracy. Discriminant analysis based on high altitude temperatures and low altitude standard meterological records is successfully applied as an alternative method of temperature extrapolation, providing a marked improvement on previous extrapolation techniques.

INTRODUCTION

As part of an investigation into an altitudinally delimited zone of erosion between 1100 and 1500 m on northerly aspects of mountains in the Wakatipu basin, Central Otago, a method for accurately estimating winter snowline elevations had to be developed. Jardine (*pers. comm.*) had proposed that the zone of erosion was related to snowline retreat caused by recent climatic warming, particularly since 1950. He postulated that the increase in the frequency of freeze/thaw cycles and needle ice erosion that accompanied this retreat could result in the depletion of the vegetative cover and soil erosion in areas previously protected by a mantle of snow during winter. To test this hypothesis past snowlines need to be determined to a high degree of accuracy. Photographic and other historical records were reviewed but were inadequate for developing a chronology of snowline changes. Consequently a computer simulation model was developed for use with standard meteorological records available from nearby weather stations.

A review of existing snow-cover models indicates that simulations of snowmelt-stream runoff relationships are most common (e.g. Anderson, 1973; Baker and Carder, 1977; Ferguson, 1984; Grimmond, 1980; Martinec, 1970). Simulations of snowline altitude have not been included in most of these models because they have been developed for hydrological modelling, where errors in snow storage and runoff caused by inaccurate estimation of snowline altitude, are assumed to be small. Methods to account for the vertical distribution of snow have included areal extent measurements derived from satellite photographs (Rango & Martinec, 1979), and zone calculations, where snow accumulation and melt are assumed to be constant within set altitudinal limits (Kuehl, 1979; Jones *et al*, 1981). Only on-site observations have been used for precise records of snowline (Green, 1975; Harrison, 1978; Moore & Owens, 1984a).

The problem of estimating the distribution of snow on maritime mid-latitude mountains, such as New Zealand's Southern Alps, has been discussed by Fitzharris (1978). He found that, in the majority of winter storms, the freezing levels and therefore the snowlines occur at an elevation above the valley floor ("rain-on-snow" conditions) and mid-winter thaw occurs frequently. Data available from the Craigieburn Range in the eastern Southern Alps indicate that 30% of annual precipitation falls as snow at altitudes above 1500 m, while snowfall is negligible below 1000 m (Morris & O'Laughlin, 1965). This situation can be contrasted with many continental areas where most winter precipitation is snow, and little or no melt of the snowpack occurs during winter. Therefore, techniques for estimating snow accumulation developed for continental areas may not be appropriate for maritime mid-latitude mountains (Fitzharris, 1978).

Archer (1970) and Fitzharris (1978) have applied simple linear and curvilinear regression models to estimate snowpack depth with altitude. Despite some good results, Fitzharris (1978) questioned the reliability of simple empirical relationships between snow accumulation and elevation in New Zealand mountains, because where "rain-on-snow" conditions occur, the increase of snow accumulation with altitude (the "snow wedge") and the snowline elevation are not consistently related to the magnitude of the snowpack. The sensitivity of the snow wedge to weather changes, from month to month and from year to year, appears to cause these inconsistencies. Fitzharris (1978) concluded that, to estimate snow pack variability, physically based models, which include variables directly related to the processes operating, should be used.

Moore and Owens (1984a) have applied a conceptual model of daily snow accumulation and ablation to a catchment in the Craigieburn Range. Extrapolating temperature and rainfall to altitude with simple linear lapse rates and using a degree-day snowmelt calculation, an adequate simulation of snow accumulation and ablation at the site of a snow course at 1750 m was made. However discrimination of rain from snow at elevations below approximately 1500 m was poor. Moore and Owens (1984a) attributed poor model performance to marked and systematic deviations of the near-surface temperature lapse rates from the assumed lapse rate of 0.65°C/100 m causing errors in extrapolating temperatures from valley weather stations.

The inadequacy of a simple linear lapse rate for extrapolating temperature with altitude, and thus determining the altitude of the rain/snow boundary, led Moore and Owens to conclude that simulations are likely to be poor unless upper level temperature data are available. Furthermore, this problem may cause errors in snowmelt calculations which also rely principally upon extrapolated air temperature data.

The aim of this study was to determine whether these problems can be overcome to obtain an adequate snowline simulation. To achieve this an alternative procedure to the simple linear lapse rate for extrapolating temperature, and thus snow accumulation and melt, is suggested. Two versions of the model are

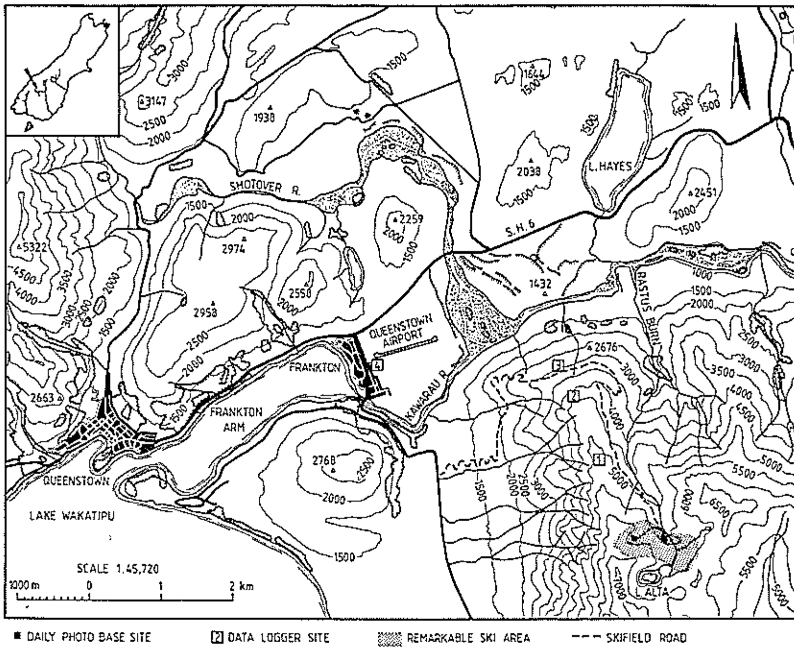


FIG. 1—Location Map.

discussed. The first uses measured temperature lapse rate data to show that snowline can be accurately modelled when temperature variation with altitude is known. The second describes a method for estimating temperature lapse rates using only meteorological data collected at valley stations. In this way long term modelling of snowlines and snow accumulation is possible without recourse to long periods of data collection at upper mountain sites.

THE STUDY AREA

The study area is approximately 50 km east of the Main Divide of the Southern Alps, South Island, New Zealand (Fig. 1). The site is on the eastern shores of Lake Wakatipu, at the northern end of the Remarkables Range. The area has experienced major faulting and folding of the underlying schist bedrock (Cox, 1985) and glacial modification by the large valley glacier that formed the Wakatipu trough. Slopes range from 20° to 50° and rock outcrops and bluffs are common on the upper slopes, particularly on the steep westerly aspect of the Remarkables which gives the range its name. Vegetation is predominantly grassland, ranging from introduced grass species and scrub on the lower slopes (400–800 m), to short tussock grassland (800–1100 m), and snow tussock grassland (1100–1600 m). Above 1600 m screes, fell field and dwarfed alpine short tussock predominate.

Mean annual precipitation is 650 mm at Queenstown Airport and 800 mm at Queenstown, some 10 km to the west. Precipitation is well distributed throughout the year. The mean seasonal snowline is approximately 1500 m, but light snowfalls frequently occur to about 1000 m, with occasional falls to the valley floor at 350 m. Snowmelt can occur throughout the winter but most melt occurs quite rapidly between September and December. By the end of December the Remarkables are normally snow free, except for a few persistent snow patches.

Mean annual temperature at Queenstown Airport is 9.7°C and at Queenstown 10.2°C. Mean daily temperatures range from 23°C in January down to between 3° and 5°C in June and July. At Queenstown Airport the mean daily minimum temperature is below freezing in June, July and August while Queenstown's mean daily minimum falls below zero only in July.

DESCRIPTION OF SNOWLINE MODEL

The modelling routines used were based upon those of Male and Gray (1981) and Moore and Owens (1984a). Snow accumulation and ablation were calculated at a series of ten-metre altitude increments using a standard degree-day melt factor. A snow wedge could then be calculated for each day from which the model could estimate snowline altitude, the lowest altitude at which snow was lying. The model requires precipitation and temperature data for each altitude increment. Precipitation data were extrapolated from Queenstown Airport using a linear lapse rate to account for increases in precipitation with altitude. Temperature lapse rates up to 1600 m were interpolated using temperature data from Queenstown Airport and three automatic stations on the Remarkables. Above 1600 m lapse rates were assumed to approximate average values for lapse rates between the two highest automatic stations (c. 05°C/100 m).

Moore and Owens (1984a) used a standard single threshold technique to differentiate rain from snow. If the estimated ambient temperature (Ta) was less than or equal to a snow/rain threshold temperature (Tb), precipitation was considered to be snow and added to the snowpack; otherwise it was treated as rain. This instantaneous transition from snow to rain is an oversimplification causing the simulated snow wedge to have a blunt tip. Typically a transition zone, where both rain and snow occur together is expected, rather than an abrupt change from rain to snow. Snow accumulation immediately above the altitude at which snow is first observed will be negligible, but as the actual freezing level is approached the proportion of precipitation falling as snow will rapidly approach 100%. The following equation (after Fitzharris, 1975), uses a double threshold to model this transition. The original Fitzharris equation used a linear transition, but the equation included here uses a parabolic transition which was found to more accurately simulate the distribution of snow accumulation described above.

$$S_{(H)} = \frac{(H - Ho)^2}{(He - Ho)^2} * P_H \qquad (1)$$

where

$S_{(H)}$ = snow deposition (mm) where $Ho < H < He$
H = altitude (m)

Ho = altitude where $Ta = Tb$
He = altitude where $Ta = 0°C$
P_H = precipitation at altitude H

Above He all precipitation was assumed to be snow, and below Ho to be all rain. Figure 2 compares the snow wedges created by the alternative functions.

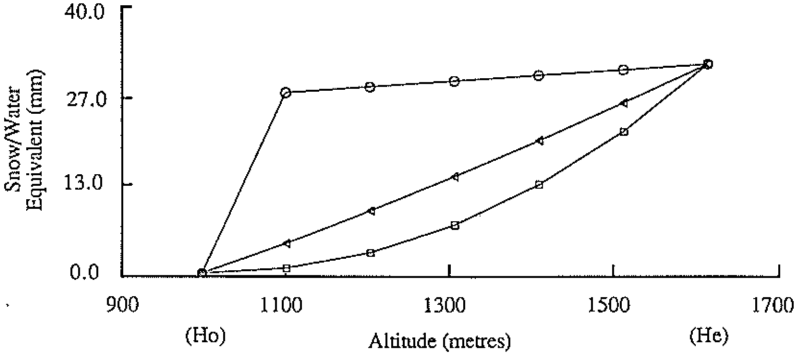


FIG. 2—A comparison of the three methods for differentiating rain from snow. The single threshold method (θ), the double threshold method (△), and the parabolic double threshold equation used in this study (⊟).

Melt occurs on days with no precipitation and temperatures greater than 0°C, or on days with precipitation and temperatures greater than Tb. For temperatures between Tb and 0°C, melt and accumulation were integrated to give net melt or accumulation according to the proportions of rain and snow at that elevation. A standard degree-day melt calculation was applied (Male & Gray, 1981, p 418; Moore & Owens, 1984a).

$$M = (MF + (0.0126 * P)) * Ta \qquad (2)$$

where

M = melt (mm/day)
P = precipitation (mm/day)
MF = melt factor (mm/day/°C)

Using this equation snowmelt is not greatly increased by rain, except during very heavy warm rainfalls. This may give conservative estimates for the New Zealand environment, where warm northwesterly rain is frequently accompanied by strong advection (Fitzharris *et al.*, 1980; Moore & Owens, 1984c); however, insufficient data are available to develop a suitable alternative equation for New Zealand conditions. Water retention capacity and refreezing coefficients were omitted because Moore and Owens (1984a) have shown they do not improve the fit of the model in a similar environment. Seasonal variations have been shown to occur in the melt factor (MF) and are approximated by applying a sine function to the degree-day melt equation, making the summer value of the seasonally adjusted melt factor (MF′) twice the winter value (Anderson, 1973).

$$MF' = MF + [MF * (0.5 * \sin(6.283 * (JD + 90)/365))] \tag{3}$$

where

JD = Julian day

DATA REQUIREMENTS

The meteorological data required for modelling were obtained from the New Zealand Meteorological Service climate station at Frankton (Queenstown Airport — NZMS station: I58074) and from three automatic stations recording air temperatures at 950, 1295 and 1615 m on the Remarkables. Upper mountain temperatures were measured using thermocouple temperature sensors and recorded by an electronic data logger. Details of instrumentation and equipment are explained in Barringer (1986). Mean temperatures calculated from daily maximum and minimum temperatures have been used in most previous studies. However, analyses of data from the Remarkables suggested that daily mean temperatures derived in this way were often not good estimates of the true daily mean temperature. This, and subsequent requirements for a new method of temperature extrapolation, led to the use of 10 am dry bulb temperatures from the three upper level sites and the 9 am dry bulb temperatures from Queenstown Airport to obtain near "instantaneous" temperature profiles. The 10 am dry bulb readings gave a consistently better estimate of daily mean temperature than did the use of maximum and minimum temperatures. Daily rainfall readings for Queenstown Airport were assumed to relate to the previous day.

Snowline altitudes were determined independently from a series of daily photographs taken from a fixed location, at approximately 9 am, conditions permitting, over the periods May to November 1984 and 1985. An overlay with marked altitude increments allowed snowline altitude ($>$ 60% snow cover) to be estimated to within ±20 m. These snowline data were used to calibrate the model.

MODEL CALIBRATION

Optimised values for model parameters were calculated by a procedure (Rosenbrook, 1960; Douglas, 1974) which simultaneously optimises all variables using a least squares criterion between the observed and modelled series to assess goodness-of-fit. The optimising program requires starting values for model parameters, as well as maximum and minimum limits, to avoid distortion of parameter values beyond values which might reasonably be expected on the basis of current knowledge. The optimiser was run several times from different starting points to prevent entrapment in local minima in the least squares response surface.

Bergstrom (1975) suggests that improvements in fit which are visible in graphical output may not be discernible in the least squares surface. Additional qualitative assessments of the sensitivity of each parameter can be made by comparing graphs of the simulated series and explaining the differences by physical reasoning (Bathurst, 1986). Physical reasoning provides extra information regarding parameter sensitivity and promotes confidence in the results. In conjunction with this final qualitative step the efficiency rating E (Nash & Sutcliffe, 1970), also derived from the sum of squares of the error, was used.

$$E = 1 - (RMSE/SD)^2 \tag{4}$$

where RMSE = the root mean square of the error between the modelled and the observed series.
SD = standard deviation of the observed series.

E may range from one to negative infinity. One represents a perfect fit, while negative values imply a poor fit, and values greater than zero indicate that the model provides a better estimate than the mean of the observed series. For a model which works reasonably well the value of E will be between zero and one. This value proved easier to interpret than a simple sum of squares value during the final graphic sensitivity analysis.

Using the method described above the optimal values for parameters were found to be:

Snow/Rain Threshold Temperature	(Tb) = 0.9°C.
All Snow Threshold Temperature	(Ts) = 0.0°C.
Precipitation Correction Factor	(PCF) = 0.8 mm/100 m.
Melt Factor	(MF) = 2.9 mm/day/°C.

The precipitation lapse rate (PCF) of 0.8 mm/100 metres is equivalent to an increase of 200% over a 1200 m range. The precipitation lapse rate will vary considerably from place to place. For example, in another study in Central Otago, an increase of 250% over a similar altitude range is quoted (Grimmond, 1980).

Other studies have shown that the melt factor varies greatly both temporally and spatially. It is difficult, however, to compare values from other studies because different temperature parameters have been used. Moore and Owens (1984a) used melt factors of 4 and 8 mm/day/°C (in conjunction with mean daily temperature) which are close to the values found by Yoshida (1962) but generally higher than those quoted elsewhere (e.g. Male & Gray, 1981).

The value for the snow/rain threshold temperature (Tb) is usually assumed to be less than 2.0°C. This can be compared with the value obtained by Crowe (1971) who found snow occurring a maximum of 300 m below the storm freezing level (assuming a lapse rate of 0.6°C/100 m this would give a maximum value of Tb $\approx$ 1.8°C). This value will vary from storm to storm and the optimal value of 0.9°C is not unreasonable given the complexity of temperature/altitude relationships. Similarly, since rain only rarely occurs at temperatures below freezing, Tb is commonly assumed to be greater than 0.0°C. Should Tb = Ts, the double threshold equation will collapse to the single threshold equation of Moore & Owens (1984a).

Figure 3a and 3b show the simulated snowline elevations for 1984 and 1985 using these optimal values and interpolated temperature lapse rates calculated from upper-mountain temperature data. Generally the modelled series provide

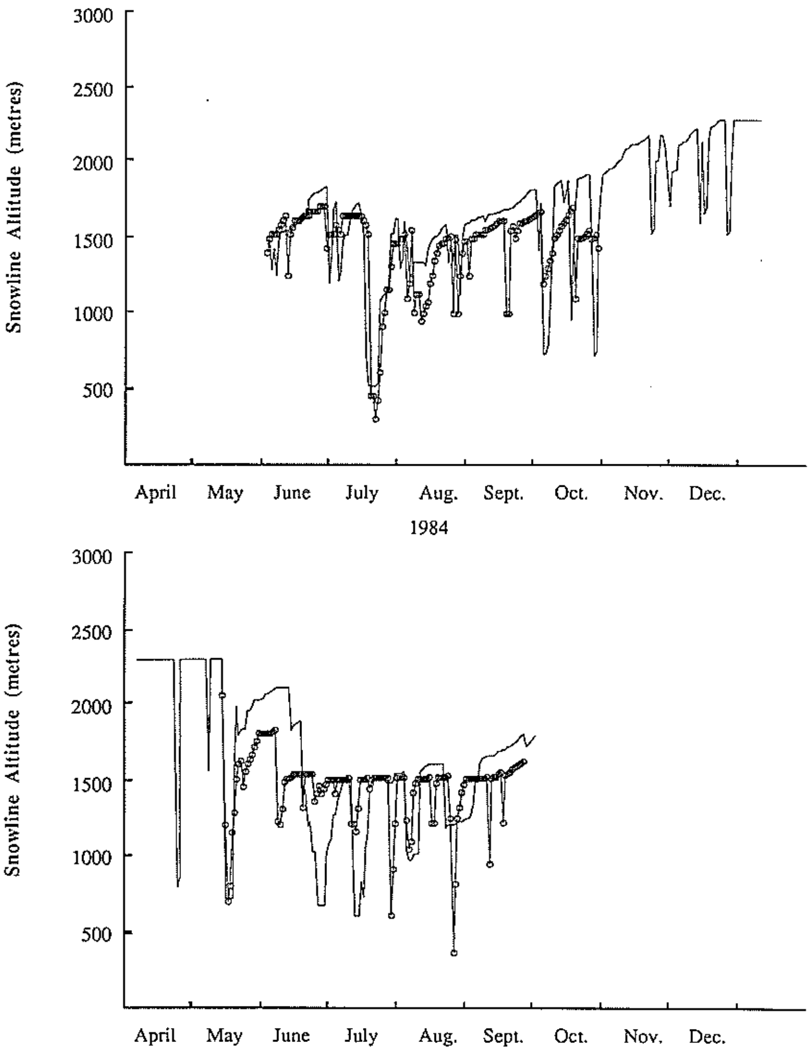


FIG. 3a & 3b—Simulated snowline data for 1984 and 1985 (—) is plotted against snowlines recorded by daily photographic survey (-o-). This simulation uses temperature data recorded on the Remarkables from May 1984 to October 1985.

good simulations of the observed series (E = 0.96 for 1984 and E = 0.80 for 1985), and the optimal parameter values all appear to be reasonable given current knowledge.

Changes in the melt factor and precipitation lapse rate have opposing effects on the simulated series, but relatively large changes in either were required before model performance was seriously affected. The snow/rain threshold temperature is important in determining the altitude of snowline during new snow events but the model appeared surprisingly robust to changes in this value. Errors due to simplification of the temperature lapse rate must occur with only four data points, but these appear insufficient to cause gross errors in the simulated snowline. This suggests that the inclusion of measured temperature lapse rates has eliminated a major source of error in previous models.

The model used in this study was calibrated only for northerly aspects. Above 1500 m no large area of northerly aspect occurs in the study area. The topography of the area results in the rate of snowline retreat above 1500 m being slower. This, and the difficulty of obtaining visual estimates of snowline, has led to a relatively poor record of real snowline above 1500 m. In fact 1500 m was close to the persistent winter snowline for other aspects, and the model may provide a better estimate of the persistent winter snowline on northerly aspects (1600 to 1700 m) than does the observed series. Nonetheless, care must be taken since the optimising procedures will attempt to override the effect of the topography by setting the melt factor too low. This results in a better fit during periods of melt but gives poorer melt response immediately after a snowfall. This was highlighted in the graphic sensitivity analysis and suggests that the most efficient fit to the observed series may not necessarily be the best representation of the snowline variations on this northerly aspect.

Bergstrom (1975) and Sorooshian (1983) indicate that at least four to five years, including a light and heavy snow year, would be desirable for calibration, since differences in meteorological conditions can cause changes in parameter optima, making the accuracy of simulations inconsistent from year to year. Only two years' data were available for this study, but despite this, the results of the 1984/5 snowline simulations indicate the model structure is essentially sound, with the simulations being relatively robust to minor changes in parameters.

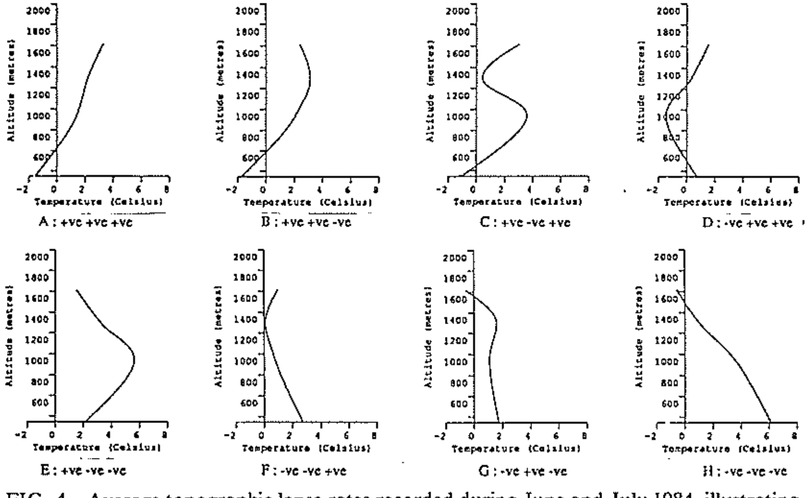


FIG. 4—Average topographic lapse rates recorded during June and July 1984, illustrating the eight types of lapse profile that may occur between the upper mountain stations and Queenstown Airport.

MODELLING USING ESTIMATED TEMPERATURE LAPSE RATES

Obled & Harder (1978) discuss the variation in temperatures and lapse rates, both with altitude and time, and point to important features such as the difference between mean lapse rates below 1500 metres (0.3 to 0.4°C/100 m) and between 1500 and 5000 metres (0.5 to 0.6°C/100 m). Similarly, while free air and near-surface temperatures are approximately equal when averaged over long periods, they may differ by as much as ±5° C for individual readings. Along with comments by Moore and Owens (1984a) regarding errors in temperature extrapolation using an assumed lapse rate of about 0.6°C/100 m, this clearly suggests that an alternative method of temperature extrapolation is required.

With both standard meteorological records from Queenstown Airport and detailed temperature lapse rate data from the Remarkables available for 1984/85 it was possible to derive, for this study area, a statistical relationship between the two, permitting the shape of the temperature lapse profile to be predicted from the Airport data.

This method was based on a discriminant analysis which was applied to the meteorological data using the temperature lapse profiles as the basis for grouping data. The groups were organised according to whether the lapse rates between the temperature stations were positive or negative. Eight possible temperature lapse profile types can be defined in this way (Fig. 4). All types of lapse profile occur naturally, but types E and H, which approximate a simple inversion and a simple negative profile respectively, are the two most common, representing about 67% of the total (Table 1). Because the proportions of each profile type

TABLE 1—A classification matrix showing the percentage of lapse rate profiles correctly classified by the discriminant analysis when using all eight profile types. Data were for the period June to August 1984.

Group	Percent Correct	Number of Cases Classified into Group								Total
		A	B	C	D	E	F	G	H	
A	25.0	1	2	0	1	0	0	0	0	4
B	83.3	0	5	0	1	0	0	0	0	6
C	71.4	0	1	5	0	1	0	0	0	7
D	100.0	0	0	0	4	0	0	0	0	4
E	72.4	0	0	6	0	21	0	0	2	29
F	60.0	0	0	0	0	1	3	0	1	4
G	100.0	0	0	0	0	0	0	2	0	2
H	54.5	2	0	1	0	3	7	2	18	33
Total	65.6	3	8	12	6	26	10	4	21	89

vary seasonally along with other standard meteorological data, particularly temperature, seasonal analyses were made.

The year was divided into four three-month seasons with winter being from June to August; spring from September to November; summer from December to February; and autumn, from March to May. The best combination of input data from Queenstown Airport was determined by the discriminant analysis to be: wind direction, dry bulb temperature, relative humidity, maximum temperature, and wind gust velocity. Table 1 shows the overall rate of correct profile discrimination to be 66%.

The rate of correct discrimination can be improved by reducing the number of grouping options, although this improvement must be offset against the associated loss of accuracy when mean temperature lapse rate is calculated for a wider range of daily profiles. Several combinations of profile groupings were tested with the best results obtained by combining the profiles of type A and B, C and D, E and G, and F and H into new groups 1, 2, 3, and 4 respectively. These were grouped on the basis of similarity of general meteorological conditions associated with profile types and on similarity of the mean profile for each group.

Correct profile discrimination with these groups averaged 65%, although some variation between seasons can be observed. Tables 2 and 3 give the results of the analysis and show the seasonal distribution of profile types.

The relative frequency of occurrence of each type of profile is shown in Table 2. In total, 63% of the period has a complete record for all three sites, and both winters have close to 100% records. No clear seasonal pattern in profile occurrence

TABLE 2—Frequency of each type of profile recorded during 1984 and 1985. Note that almost all data loss occurred during spring, summer and autumn. Winter records are close to complete.

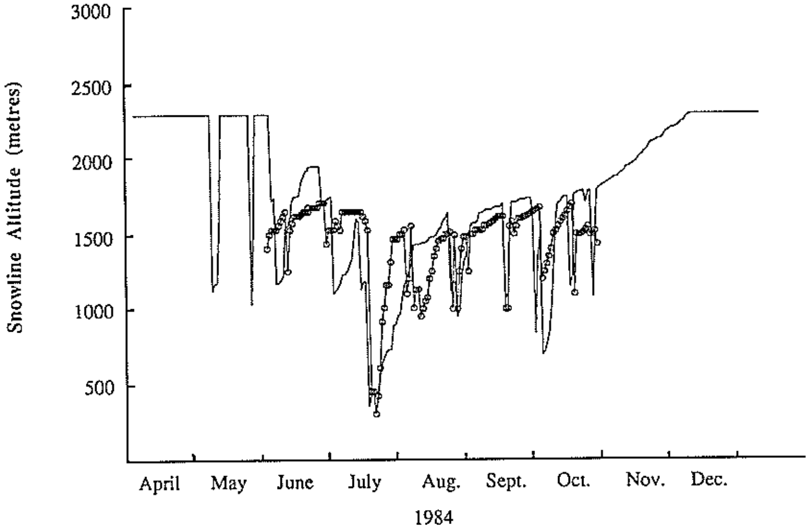
TABLE 3—Percentages of correct profile discrimination using five input variables (wind direction, dry bulb temperature, relative humidity, maximum temperature and wind gust velocity).

Group	Total Number of Profiles				
	1984			1985	
	Winter	Spring	Summer	Autumn	Winter
1	10	2	6	5	10
2	11	1	8	6	12
3	30	19	16	5	28
4	39	18	16	6	28
Total	90	40	46	22	85

Group	Percent Correct				
	1984			1985	
	Winter	Spring	Summer	Autumn	Winter
1	70.0	100.0	33.3	80.0	80.0
2	72.7	100.0	62.5	50.0	41.7
3	80.0	63.2	50.0	80.0	50.0
4	79.5	83.3	62.5	100.0	31.4
Total	77.8	75.0	54.3	77.3	44.7

is apparent. Spring 1984 has a low proportion of complex type 1 and 2 profiles, but in all other seasons (except autumn 1985 which is affected by data loss) there is a dominance of types 3 and 4 in an approximate ratio of 2 to 1.

Table 3 demonstrates the instability of the discriminant analysis with summer and winter 1985 having significantly lower rates of discrimination than other



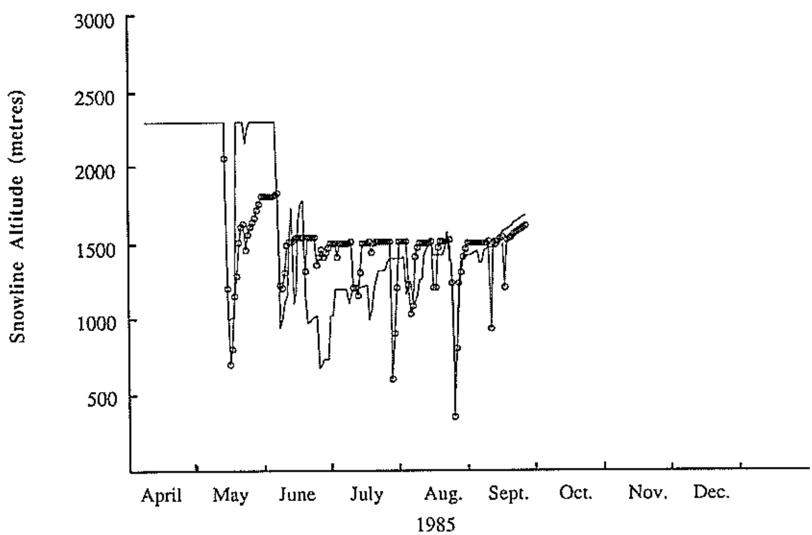


FIG. 5a & 5b—Simulated snowline data for 1984 and 1985 (—) is plotted against snowlines recorded by daily photographic survey (-o-). The simulation uses temperature data estimated using the discriminant analysis method described.

seasons. Winter 1985 in particular had poor results. It is not entirely clear why this should have been the case. It does not appear to be a function of the amount of missing data because winter 1985 has a complete record while the summer does not. There may be some relationship to the stability of the local synoptic conditions during the period. For example, although the mean barometric pressure was approximately the same during winter 1984 and 1985, it was lower during the occurrence of type 1 and 2 lapse profiles in 1985 by an average of 7 millibars. Such differences may have been sufficient to disrupt the discriminant analysis. Whatever the cause of this problem, the overall rates of correct profile discrimination were sufficiently high to suggest that this method offers a practical alternative to the fixed negative lapse rate approach.

Using the discriminant functions derived from the analysis, average lapse rates for the four temperature lapse profile groups were calculated for winter, spring, summer and autumn. Meteorological data were then processed through the discriminant functions so that temperature at any altitude could be estimated using the temperature at the base of the mountain and the appropriate average lapse rate.

The optimal variable values obtained using discriminant analysis were:

melt factor	(MF) = 3.6 mm/day/°C.
snow/rain threshold temperature	(TB) = 0.65°C.
all snow threshold temperature	(TS) = 0.0°C.
precipitation correction factor	(SCF) = 0.8 mm/100 m.
Snow correction factor	(SCF) = 1.3.

These values gave a model efficiency rating E = 0.96 for 1984 and E = 0.68 for 1985. The plots of simulated versus real snowline altitude (Figs. 5a and 5b) indicate that the model provides a good estimate of snowline altitude below 1500 m during midwinter. The persistent overestimation of snowline altitude above 1500 m apparent when using upper mountain temperatures remains, and appears to be the only major error.

CONCLUSIONS

The failure of most snow models to reproduce variations in snow accumulation and ablation with altitude limits their ability to model snow-related catchment hydrology accurately. This stems from the inaccuracy of simple linear lapse rates for characterising temperature variations with altitude during snowfall and melt periods. In this study the use of measured temperature lapse rates provided a marked improvement to the simulation of snowline altitude. A persistent overestimation of snowline altitude during long melt periods remains the major source of error in the simulations. This appears to be a function of the study area's topography, a factor which the model was not designed to simulate, rather than an inherent fault in basic snow accumulation and melt calculations. The use of discriminant analysis for estimating temperature lapse rates from standard meteorological data from a valley site has proven to be a viable alternative to actual measurements of temperature lapse rates. As a result it should be possible, after a calibration period, to model current snow accumulation and ablation without requiring expensive long-term temperature recording. It should also be possible to reconstruct past snowline data when upper mountain temperature data were not available.

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REFERENCES

Anderson, E. A., 1973: National Weather Service River Forecast System — Snow accumulation and ablation model. *NOAA — Technical Memorandum NWS — Hydro-17:* 219 p.

Archer, A. C., 1970: Studies of Snow Characteristics in the Northeastern Ben Ohau Mountains, New Zealand. *Journal of Hydrology (N.Z.)* 9: 4–21.

Baker, M. B.; Carder, D. R., 1977: Comparative Evaluations of Four Snow Models. *Proceedings of the 45th Annual Meeting of the Western Snow Conference, Albuquerque, New Mexico:* 58–62.

Barringer, J. R. F., 1986: Soil Erosion in relation to Snowline in the Remarkables, Central Otago, unpub. MSc thesis, Geography Department, University of Otago, New Zealand.

Bathurst, J. C., 1986: Sensitivity of the Systeme Hydrologique Europeen for an Upland Catchment. *Journal of Hydrology* 87: 103–123.

Bergstrom, S., 1975: The Development of a Snow Routine for the HBV-2 Model. *Nordic Hydrology* 6: 73–92.

Cox, S., 1985: Remarkables Geology, unpub., BSc (hons) dissertation, Department of Geology, University of Otago, New Zealand.

Crowe, P. R., 1971: *Concepts in Climatology*, Longman Group, London.

Douglas, J. R., 1974: Conceptual Modelling in Hydrology. *Natural Environmental Research Council, Institute of Hydrology Report No. 24:* 2–29.

Ferguson, R. I., 1984: Magnitude and Modelling of Snowmelt Runoff in the Cairngorm Mountains, Scotland. *Hydrological Sciences Journal* 29(1): 49–62.

Fitzharris, B. B., 1975: Snow Accumulation and Deposition in on a West Coast Mid-Latitude Mountain, unpub., Ph.D. thesis, University of British Columbia.

Fitzharris, B. B., 1978: Problems in Estimating Snow Accumulation with Elevation on New Zealand Mountains. *Journal of Hydrology (N.Z.)* 17(2): 78–90.

Fitzharris, B. B.; Stewart, D.; Harrison, W., 1980: Contribution of Snowmelt to the October 1978 flood of the Pomohaka and Fraser Rivers, Otago. *Journal of Hydrology (N.Z.)* 19: 84–93.

Green, F. H. W., 1975: The Transient Snowline in the Scottish Highlands. *Weather* 30(7): 226–235.

Grimmond, C. S. B., 1980: Runoff from Seasonal Snow in the Fraser Catchment, Central Otago, unpub., BSc (hons) dissertation, Geography Department, University of Otago, New Zealand.

Jones, E. B. *et al.*, 1981: Application of a Snowmelt Model to Two Drainage basins in Colorado. *Proceedings of the 49th Annual Meeting of the Western Snow Conference, St. George, Utah:* 43–54.

Kuehl, D. W., 1979: Volume Forecasts Using the SSARR Model in a Zone Mode.

Proceedings of the 47th Annual Meeting of the Western Snow Conference, Sparks, Nevada: 38–47.
Male, D. M.; Gray, D. M., 1981: *Handbook of Snow; Principles, Processes, Management and Use*, Pergamon Press, Toronto, 776 p.
Martinec, J, 1970: Study of Snowmelt Runoff Process in Two Representative Watersheds with Different Elevation Range. *IAHS — UNESCO Symposium on the Results of Research on Representative and Experimental Basins, Wellington, New Zealand, International Association of Hydrological Sciences Publication No. 96:* 29–39.
Moore, R. D.; Owens, I. F., 1984a: Modelling Alpine Snow Accumulation and Ablation Using Daily Climate Observations. *Journal of Hydrology (N.Z.)* 23(2): 73–83.
Moore, R. D.; Owens, I. F., 1984b: A Conceptual Runoff Model for a Mountainous Rain-on-Snow Environment, Craigieburn Range, New Zealand. *Journal of Hydrology (N.Z.)* 23(2): 84–99.
Moore, R. D.; Owens, I. F., 1984c: Controls on Advective Snowmelt in a Maritime Alpine Basin. *Journal of Climate and Applied Meteorology*, 23: 135–142.
Morris, J. Y.; O'Loughlin, C. L., 1965: Snow Investigations in the Craigieburn Range. *Journal of Hydrology (N.Z.)* 4(1): 2–16.
Nash, J. E.; Sutcliffe, J. V., 1970: River Flow Forecasting through Conceptual Models, Part 1: A discussion of principles. *Journal of Hydrology* 10: 282–290.
Obled, Ch.; Harder, H., 1979: A Review of Snowmelt in the Mountain Environment. *Proceedings, Modelling of Snow Cover Runoff*, Colbeck, S. C. & Ray, M. (eds). U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, pp 179–204.
Rango, A.; Martinec, J., 1979: Application of a Snowmelt — Runoff Model using Landsat Data. *Nordic Hydrology* 10: 225–238.
Rosenbrock, H. H., 1960: An Automatic Method of Finding the Greatest or Least Value of a Function. *Computer Journal* 3: 175–184.
Sorooshian, S., 1983: Surface Water Hydrology: on-line estimation. *Reviews of Geophysics and Space Physics* 21: 706–721.
Yoshida, S., 1962: Hydrometeorological Study on Snowmelt. *Journal of Meteorological Research* 14: 879–899.