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PROBLEMS IN ESTIMATING SNOW ACCUMULATION WITH ELEVATION ON NEW ZEALAND MOUNTAINS

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ABSTRACT

Freezing level and hence snowline intersect New Zealand mountains above their bases during most winter storms. On two North American and two New Zealand mountains the snow pack increases with elevation as a snow wedge. The shape of the wedge is not simple, changing inter- and intraseasonally. This renders the traditional single snow course ineffective in estimating total areal snow pack in New Zealand. The varying shape of the snow wedge also means snow line data alone are not sufficient to estimate water stored as snow. Empirical methods are unlikely to produce reliable predictions; a more theoretical approach considering factors controlling the shape of the snow wedge is needed.

INTRODUCTION

This paper examines the possibility of establishing workable empirical relationships between seasonal snow accumulation and elevation on New Zealand mountains. These relationships would be of the form $w(z, t) = f(z, zo)$, where w is water equivalent of the snow pack at some elevation z, and zo is the elevation of the seasonal snow line at time t. The utility of the snow course concept for indexing snow accumulation on New Zealand mountains is related to this question, and is also examined. Accumulation is here used to mean the balance of the snow (water equivalent) that remains on the ground after melt and sublimation.

A New Zealand mountain may be more broadly defined as a *west coast midlatitude mountain*, that is one on which the freezing levels, and hence the snow lines, during most winter storms intersect the mountain at some elevation above its base. Such mountains are located in the maritime climate of the disturbed westerlies (e.g. New Zealand, British Columbia, Washington, Oregon, Chile). Mountainous river catchments within these areas are sometimes called 'rain–on–snow basins' (Rockwood and Anderson, 1970). It is important to distinguish a west coast midlatitude mountain from that located in a continental situation, where the freezing level for much of the winter is at, or below, the base of the mountain. Then most winter precipitation falls as snow, and there is little

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midwinter melt of the snow pack. Therefore techniques for estimating snow accumulation developed for continental areas may not be appropriate on west coast midlatitude mountains.

In New Zealand, the study of seasonal snow accumulation with elevation is of some relevance. On average, snow melt provides 10–25% of annual runoff of South Island mountain rivers, but its contribution is highly variable from year to year (Fitzharris, in press). Good estimates of snow accumulation on mountains would allow improved management decisions about the control of this snow melt runoff for power generation, irrigation and flood protection. These estimates would also be useful for the understanding of the local ecology, for feasibility studies of new ski areas, and for the prediction of avalanches, snow loads on roofs, and snow creep pressures against structures such as transmission towers.

EXAMPLES CHOSEN FOR ANALYSIS

Four examples, two from North America, and two from New Zealand, are employed to illustrate snow accumulation with elevation (Fig. 1). The Mount Seymour observations were made by the author. The observations for the Willamette Basin are reported in U.S. Army (1956), for the Tasman Glacier in Ministry of Works (1968), and for the Ben Ohau Range in Archer (1970). Free air freezing levels from nearby radiosonde stations suggest that snow lines from most winter storms do intersect these mountains (Table 1), although other properties of each example differ in detail (Table 2).

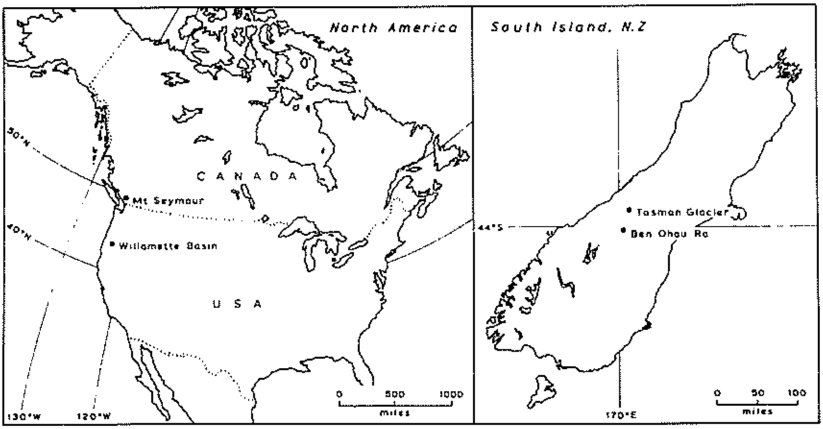

FIG. — 1 Location of mountains chosen as examples.

The Willamette Basin data are included because they represent extensive early work on the type of mountain of concern here. The mountains of the Basin differ from those in New Zealand in that the zone of seasonal snow cover is forested. Mount Seymour is also forested, although the trees thin markedly at the higher elevations. On this mountain the water equivalent of the snow pack was measured at about 100 m elevation intervals with a Federal snow sampler. All measurement sites were within open areas greater than three tree heights in diameter, and were relatively homogeneous in slope and aspect. The snow cover on the Tasman Glacier may not be typical of mountain slopes, but these data are

TABLE 1 — Mean altitude (geopotential metres) of winter freezing levels in the free atmosphere near the mountains chosen as examples.

Month	*Invercargill (N.Z.)*	*Tatoosh Is (Olympic Pen., U.S.A.)*	*Month*
Apr	2319	2440	Oct
May	1876	1730	Nov
Jun	1635	1200	Dec
July	1514	860	Jan
Aug	1545	1060	Feb
Sept	1666	1020	Mar
Oct	2078	1420	Apr
Nov	2202	2240	May

Source: Tatoosh Is. data are for period 1946-55 and were computed from Ratner (1957). Invercargill data from "Summaries of radiosonde data, 1956-61", *N.Z. Met. Serv., Misc. Pub. 119.*

included because of the great range of altitude, up to and beyond the glacier equilibrium line.

INFLUENCE OF VARIABLES OTHER THAN ELEVATION

Apart from elevation, other variables exert important influences on snow accumulation. Differences of 2% to 138% have been reported with aspect (Meiman, 1968). Packer (1962, Idaho) shows a linear increase of 0.01 m per 10% decrease in forest canopy, and Kittredge (1953, California) gives values from 0.01 to 0.06 m per 10% change in crown density for various tree species. In another study, Anderson and Pagenhart (1957, California) showed, aside from elevation, that solar energy received, and vegetation were important parameters. Hence the data collected from Mount Seymour on relatively constant aspects and slopes, and that from vegetation free Tasman Glacier are particularly valuable. For some locations, distance from the sea is likely to be an additional influence. Variables such as vegetation, slope, and aspect, tend to be most important at the local scale.

When considering either terrain segments that are homogenous, or medium to large mountain catchments where local effects average out, elevation becomes a major influence. Increases of snow accumulation of 0.1 m per 100 m increase in elevation are not uncommon. Mixsell *et al.,* (1951) report increases of 0.18 m/100 m for the Sierra Nevada of California. In Colorado 0.22 m/100 m was measured by the U.S. Soil Conservation Service (1965-67). The U.S. Army (1956) gives a rate of change of snow accumulation with elevation in the Willamette Basin of 0.25 m/100 m. On the other hand, some continental, low precipitation regions display smaller gradients. For example, Gary and Coltharp (1967) give 0.05 m/100 m for New Mexico, and Packer (1962) 0.09 m/100 m for Idaho.

TABLE 2 — Characteristics of mountain areas chosen as examples

Characteristic	*Mount Seymour*	*Willamette Basin*	*Ben Ohau Range*	*Tasman Glacier*
Country	B.C. Canada	Oregon, U.S.A.	South Is., N.Z.	South Is., N.Z.
Slope range (degrees)	8	20	15	12
Aspect range (degrees)	25	225	75	75
Elevation range (metres)	1260	853	671	1320
Max. elevation sampled (metres)	1260	1500	1646	2340
Max. snow water equivalent sampled (cm)	236	203	40	596
Vegetation	hemlock, douglas fir cedar	douglas fir, hemlock noble fir	snow tussock, alpine scrub	none (ice)
Years of observation	1969-71	1949-51	1966	1968-69
Reference	author's observations	U.S. Army (1956)	Archer (1970)	Ministry of Works (1968)

ANALYSIS OF DATA

Snow water equivalents for the four examples were first graphed against time for various elevations up the mountain. The water equivalent for the first day of each month was read from these graphs and plotted as a function of elevation. Only some of the resultant plots of w *(z, t)* are shown here (Fig. 2). Similar plots for the Willamette Basin (U.S. Army, 1956, Fig. 1, Plate 3.3) suggested that water equivalent (the ordinate) increased as a simple linear function with elevation (the abscissa) from the snow line. The line representing this function, together with the abscissa, defined a *snow wedge.* However, the snow wedges produced on Mount Seymour and on the Tasman Glacier do not always show constant increases of water equivalent with elevation (Fig. 2).

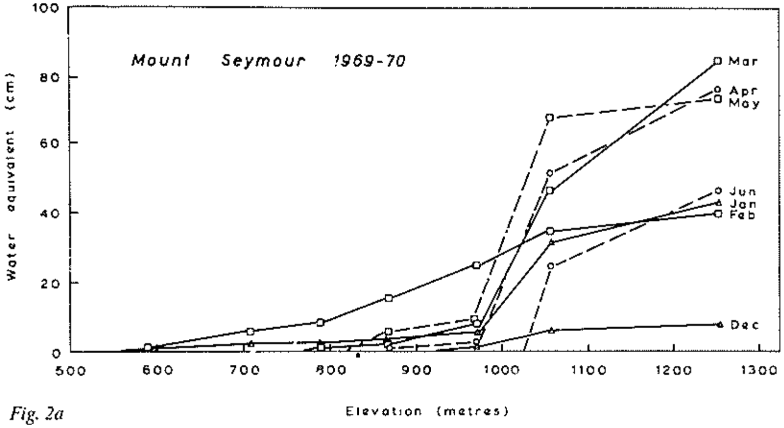

Fig. 2a

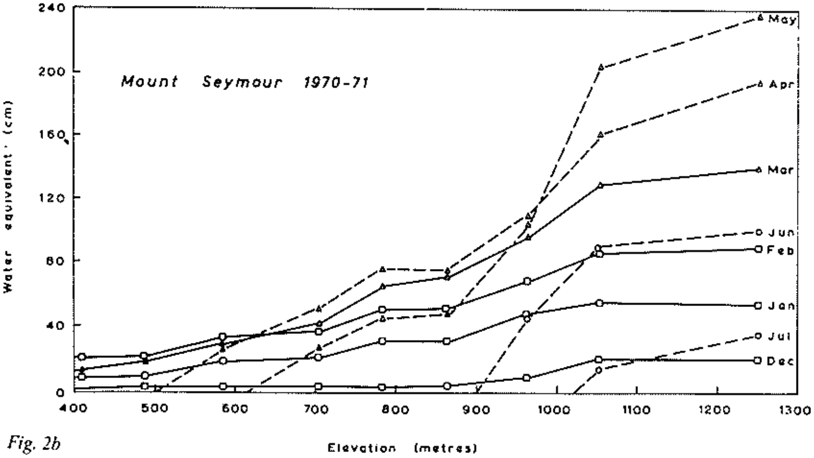

Fig. 2b

FIG.— 2a,b,c. Variations of snow pack water equivalent with elevation on the first day of each month. Tasman Glacier example compiled from data in N.Z. Hydrology Annual (1968).

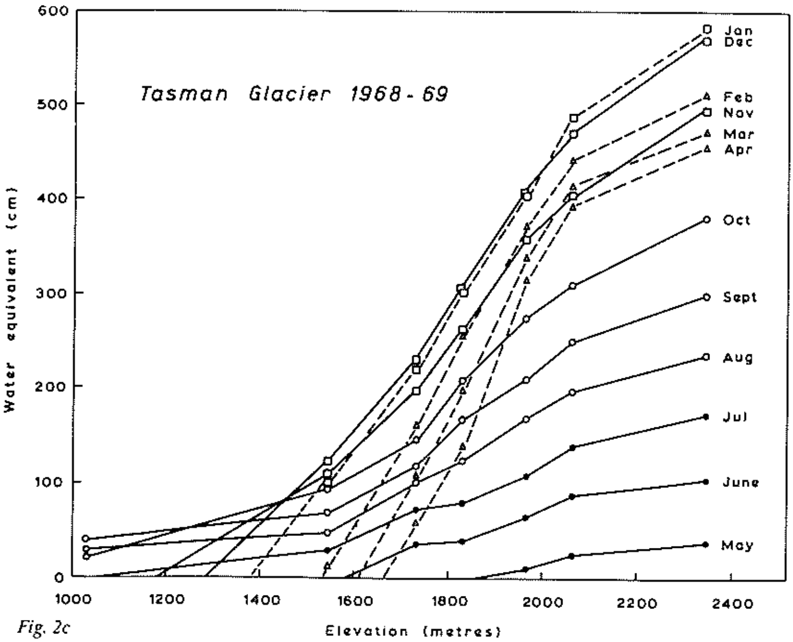

Fig. 2c

Various linear and curvilinear functions were fitted to the upper surfaces of the Mount Seymour snow wedges (Table 3). Straight line and logarithmic functions were fitted by least squares techniques and gave coefficients of determination (r^2) greater than 0.90. Inverse hyperbolic functions were fitted by trial and error. For early winter months, straight line fits proved best, but as the season progressed, curves such as inverse hyperbolic sine functions were more appropriate. Finally, during spring melt the increase of snow accumulation with elevation was well described by logarithmic functions.

If during a winter, freezing levels are maintained close to the base of the mountain, allowing little melt or rain, then there is no opportunity to develop 'steps' in the snow wedge. In these cases, it would appear that the period for which a straight line function is appropriate for describing the increase of snow accumulation with elevation can be extended. This occurred in the cool winter of 1970–71, when a straight line function provided good fits to the snow wedge up to April. On the other hand, the Mount Seymour data for the mild winter of 1969–70 allowed a linear fit only in December and February. Curvilinearity quickly developed in the snow wedge because of frequent melt periods and little snow below 900 m.

Despite these generalisations, the empirical functions developed in Table 3 are of little predictive value for estimating the increase of snow accumulation with elevation. The form of the functions and their coefficients do not behave consistently from month to month or year to year. An analyst would not know

TABLE 3 – Functions fitted to snow wedges of Mount Seymour for each month of winters 1969–70, 1970–71.

w = water equivalent of snow pack (10^{-2}m), z = elevation (m).

Month	*Winter 1969–70*	*Winter 1970–71*
Dec	$w = -20 + 0.02\,z$	$w = -9 + 0.02\,z$
Jan	$w = 23 \pm 11 \sinh^{-1} \left\lvert \frac{z-1030}{80} \right\rvert$	$w = -18 + 0.05\,z$
Feb	$w = -36 + 0.06\,z$	$w = 25 + 0.10\,z$
Mar	$w = 26 \pm 25 \sinh^{-1} \left\lvert \frac{z-1020}{40} \right\rvert$	$w = -68 + 0.17\,z$
Apr	$w = 25 \pm 13 \sinh^{-1} \left\lvert \frac{z-1020}{10} \right\rvert$	$w = -129 + 0.26\,z$
May	$w = 40 \pm 7.5 \sinh^{-1} \left\lvert \frac{z-1020}{3.4} \right\rvert$	$w = 120 + 23 \sinh^{-1} \left\lvert \frac{z-970}{5} \right\rvert$
June	$w = -1340 \pm 195 \ln z$	$w = -1931 + 287 \ln z$
July	$w = 0$	$w = -975 + 142 \ln z$

when in the season to change from a linear to a curvilinear equation. His choice of coefficients would be arbitrary, and likely to be wrong. For example, the intercept values for the straight line relationships of Table 3 vary between –0.09 m and –1.29 m water equivalent, and differ for the same month from one winter to the next. Even the mean slope of the snow wedge changes, generally becoming larger during the accumulation season, and remaining constant or decreasing during the melt period (Fig. 3). Moreover, the Mount Seymour data illustrate that the behaviour of this mean slope need not be the same from year to year.

The U.S. Army (1956) believed that during the accumulation season in the Willamette Basin there existed a simple linear relationship between the mean slope of the snow wedge and water equivalent at the upper sample limit of the mountain (Fig. 4). This implied that a single monthly measurement of snow pack water equivalent at the top of the mountain would allow definition of the function $w(z, t)$. Unfortunately, this simple relationship does not appear to be valid for other west coast midlatitude mountains. Further, for the Mount Seymour examples the relationship is different in successive years (Fig. 4). The failure of this method to satisfactorily estimate $w(z, t)$ becomes even more apparent when it is remembered that the upper surface of the snow wedge is not always a straight line, as implied by the use of the term 'mean slope'. A similar analysis of the Mount Seymour data, but for homogeneous sections of the snow wedge, displayed some trends, but these were still not consistent from section to section or year to year (Fig. 5).

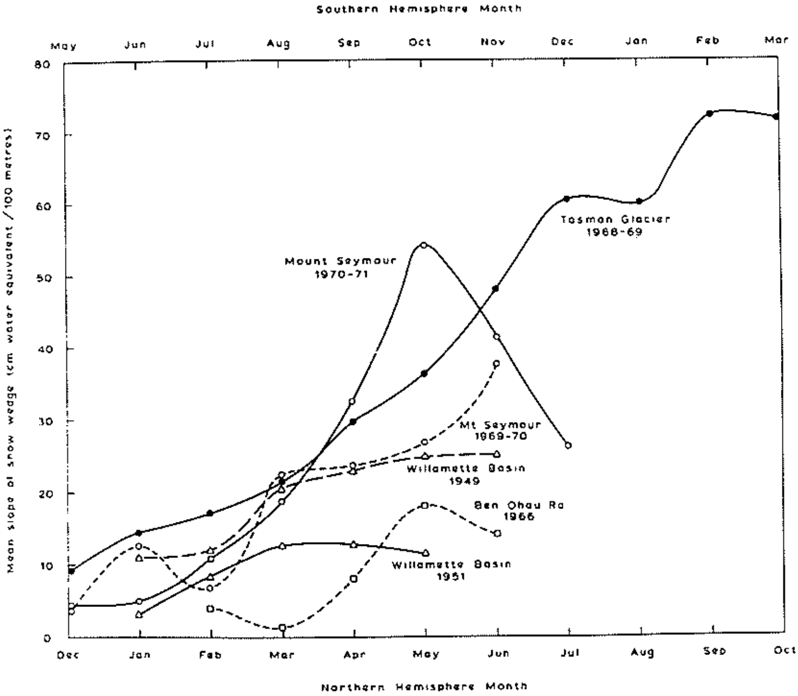

FIG. – 3 Seasonal variation in mean slope of snow wedge.

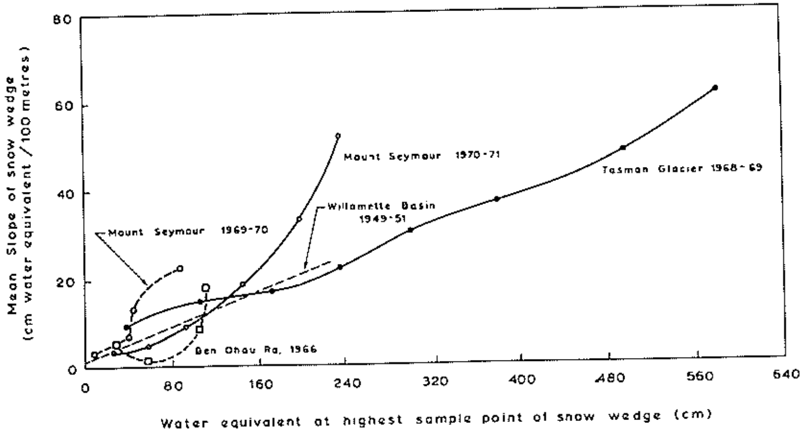

FIG. – 4 Mean slope of snow wedge during the accumulation period as a function of water equivalent at the upper sampling limit on the mountain.

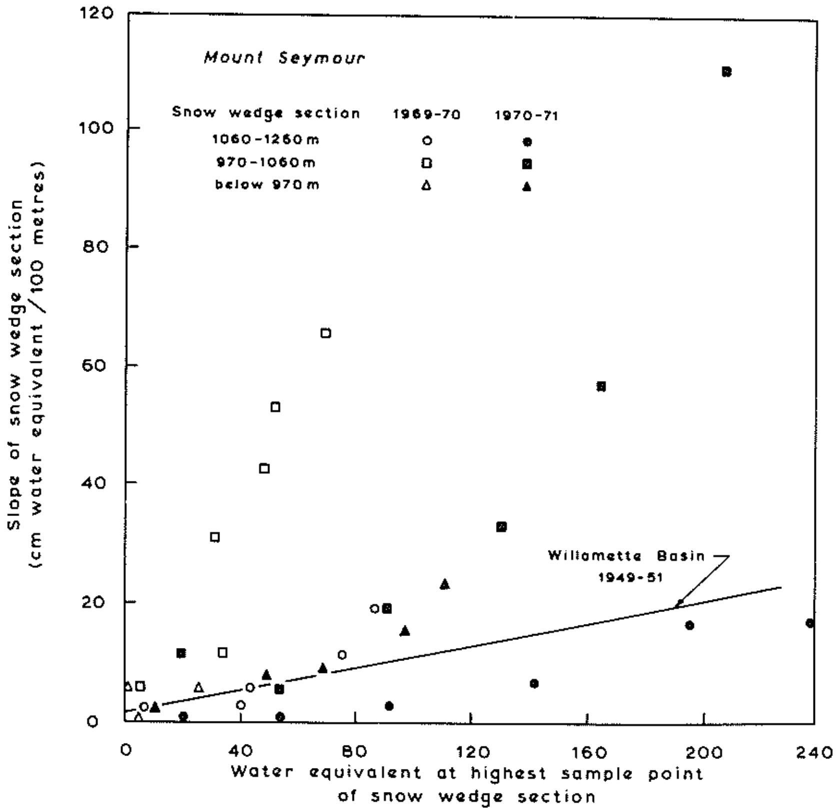

FIG. — 5 Slope of sections of snow wedge during the accumulation period as a function of water equival at the upper limit of each section.

There is also no apparent relationship between the water equivalent of the snow pack at the upper sampling limit of the snow wedge and the position of the snowline (Fig. 6). While the points on the graph describe a type of elliptical path through the season (e.g. Mount Seymour, winter 1970–71), this path is not identical from year to year.

SOME IMPLICATIONS

The previous discussion showed that on west coast midlatitude mountains like those in New Zealand, snow accumulation increases with elevation in the form of a snow wedge. However, the shape, slope and relationship with the snow line of this snow wedge is not consistent from month to month, year to year or locality to locality. It must then be concluded that empirical relationships of the form *w (z, t)* will not produce reliable predictive models on this type of mountain. The approach fails because the mean slope of the snow wedge and snow line elevation are not consistently related to the magnitude of the snow pack at a particular elevation, nor to time of year.

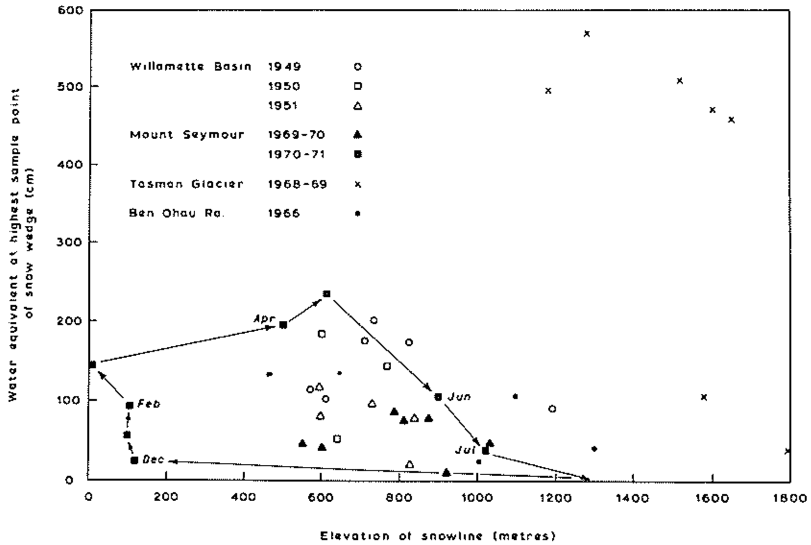

FIG. — 6 Water equivalent at upper sampling limit of snow wedge as a function of elevation of the snow line.

On west coast midlatitude mountains it is possible for water equivalents to be similar in different years (or months) at one elevation, but the distribution of water equivalent with elevation to be markedly different from year to year. A hypothetical example is shown in Fig. 7. This sharp elevation–time interaction effect, also displayed in the earlier examples, has been reported elsewhere (Packer, 1962; Anderson and West, 1965). The interaction implies that a single snow course at one elevation cannot provide a good index of snow accumulation on a New Zealand mountain.

For the hypothetical example, a snow course at elevation 1300 m records the same water equivalent in year 1 as in year 2. Yet the volume of water stored as seasonal snow in year 1 is 26 percent greater than in year 2. By relating the shape of the snow wedge to the hypsometric curve, it is apparent that the water equivalent of the snow pack at the highest elevation sampled need not be indicative of the total amount of water stored on the mountain. Further, samples of the shallow snow pack near the seasonal snow line assume greater importance, because when combined with the relatively large area of any catchment at this elevation, they are shown to produce a significant contribution to the total water stored as snow.

It also follows that knowledge of the position of the snow line, and hence area covered by snow, does not, alone, provide sufficient information to assess water stored as snow within a New Zealand catchment. Therefore satellite photographs of mountain snow cover need to be combined with ground observations to elicit the shape of the snow wedge above the snow line, before useful results for runoff prediction can be obtained.

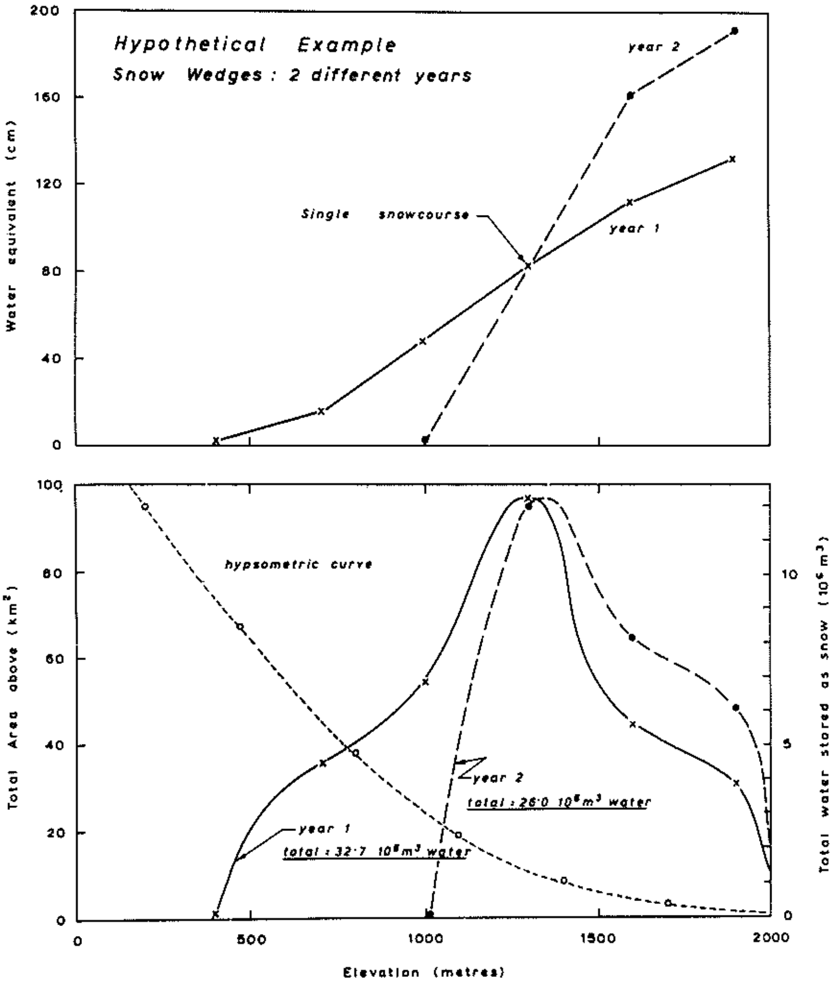

FIG. — 7 Hypothetical example of hypsometric curve, snow wedges and total water stored as snow in two different years.

DISCUSSION AND CONCLUSION

Simple empirical relationships between snow accumulation and elevation are not likely to be reliable on New Zealand mountains. This is a consequence of the formation of a snow wedge whose shape and slope appear to be very sensitive to weather changes from month to month, and winter to winter. These findings indicate that a successful approach to estimation of snow pack variations with elevation should be more physically based by including variables that are directly related to the processes operating. From detailed examination of the behaviour of snow wedges on mountains, both here and in North America, it is suggested that three critical weather factors influence their shape. They are: the variation of precipitation with elevation; the variation of melt with elevation; and the elevation of rain/snow boundary during successive storms. Models to predict snow wedge behaviour should simulate on a day–to–day basis these factors. Such an approach has been successfully applied to the Willamette Basin data and should be appropriate to New Zealand (SSARR snow melt watershed model, Rockwood and Anderson, 1970).

Since the shape of the snow wedge is particularly sensitive to the position of the rain/snow boundary from successive storms, a climatological inventory of storm types, freezing levels, and snow lines for each winter would also be useful.

It is significant that in those continental mountain regions where the freezing level of winter storms is at (or below) the base of the mountain, the elevation–time interaction effect does not exist. Consistent empirical relations describing $w(z, t)$ are then possible (e.g. in Western Colorado, U.S. Soil Conservation Science, 1965–67).

On New Zealand mountains, the changing shape of the snow wedge, and resultant elevation–time interaction effects, render unsuitable the traditional snow course where measurements are taken at a single elevation. More reliable indices of snow accumulation will be obtained with a series of snow courses at different elevations on a sample mountain, and with observations of the snow line, possibly obtained from aerial observation or satellite photography.

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