

Simulations of seasonal snow for the South Island, New Zealand

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Abstract

Seasonal snow simulations are produced for the South Island of New Zealand using a relatively simple temperature-index snow model. Results show that the snow simulations are broadly consistent with the observed snow climatology, especially with respect to estimates of snow volume and snow duration. For the parameter sets tested, snow simulations were more realistic for those for which there is strong seasonal variability in melt and where the temperature threshold for snow accumulation is set to 1°C. However, we find strong interactions among the different parameters in the snow model that cannot be resolved with the available snow data. In the short-term, detailed basin-specific studies are necessary to refine model parameter values.

Introduction

Interannual variability in seasonal snow directly affects New Zealand's economy. Snow contributes up to 24% of the annual inflows to the major lakes (McKerchar *et al.*, 1998), and many agricultural applications rely on the snowmelt component of streamflow for

their operations. Snow provides numerous opportunities for winter recreation: for example, in 2005 the five ski areas in the Southern Lakes region registered 625,198 skier days during the winter season, of which 578,308 are estimated to come from outside the region (NZTRI, 2005). Snow is also a hazard. Every year, avalanches cause the Milford Road to be closed (Hendrikx *et al.*, 2005) and heavy snow at low elevations often results in losses of livestock and damage to buildings and infrastructure (Hendrikx, 2007). The year-to-year variability in the amount of seasonal snow creates uncertainty in the energy (hydro-power generation), agriculture (availability of water for irrigation), and tourism (winter recreation) sectors.

Seasonal snow cover is also a sensitive indicator of climate change. The spatial extent of snow in the Northern Hemisphere has decreased over recent decades (Robinson and Frei, 2000; Lemke *et al.*, 2007), and regional trends in snow are quite pronounced – in western North America, for example, there have been substantial decreases in the water equivalent of the mountain snowpack accompanied by a shift to earlier timing of

snowmelt (Regonda *et al.*, 2005; McCabe and Clark, 2005; Mote *et al.*, 2005). Such trends in snow cover are closely linked to trends in temperature, and some authors speculate that decreases in snow storage are attributable to enhanced atmospheric concentrations of greenhouse gases (e.g., Mote *et al.*, 2005). Meanwhile, little is currently known about the long-term variability of snow cover in the Southern Hemisphere (Lemke *et al.*, 2007).

Because of the current lack of systematic snow observations in New Zealand (Fitzharris *et al.*, 1999), the only reasonable way to generate information regarding interannual variability and trends in seasonal snow storage is to use physically-based simulation models. Simulations of snow in New Zealand have previously been produced using the *SnowSim* conceptual snow model (e.g., Fitzharris and Garr, 1995; Fitzharris, 2004; Kerr, 2005). Fitzharris and Garr (1995) demonstrate that the *SnowSim* model matches point snow observations in the Pukaki catchment reasonably well. Similarly, Kerr (2005) shows reasonable agreement between the *SnowSim* model and point snow observations and lake inflows in the Pukaki catchment.

Given the economic importance of snow in New Zealand, it is important to assess whether the *SnowSim* model can be improved. Melt in *SnowSim* is parameterized as an empirical function of temperature, and like other conceptual snow models (e.g., Hock, 2003), *SnowSim* does not include factors known to be important in New Zealand, such as spatial variability in snow water equivalent within a model element (Liston, 2004), and accelerated melt during warm, humid, windy conditions when turbulent heat transfers of sensible and latent heat are both directed toward the snow surface and dominate the energy budget (Owens and Fitzharris, 2004). While it is possible to build more complex snow models (e.g., Slater *et al.*, 2001; Etchevers *et al.*, 2004), such models typically require additional data (e.g., wind,

humidity, radiation). Further, increased model complexity does not necessarily translate into improved model performance, particularly when data to parameterize the model are sparse (e.g., Kirchner, 2006). A key question then is to determine what level of model complexity can be supported by the available data.

The goal of our investigation is to produce credible simulations of the snow climatology for the South Island of New Zealand. This is a necessary first step in our snow modeling programme, and future work will consider simulations of the interannual variability of the snow resource and provide a detailed evaluation of snow simulations in individual basins. In this paper we seek to achieve our goal by using a simple temperature-index snow model, deliberately designed to have the same level of complexity (and include the same processes) as the *SnowSim* model. A brief description of our snow model and a comparison with *SnowSim* are provided in Section 2, the model configuration and input data are described in Section 3, the data for model evaluation are discussed in Section 4, and the model simulation results are presented in Section 5. The paper concludes with a discussion of potential areas for model improvement.

Snow model

Basic snow model

A very simple and parsimonious snow model is

$$\frac{dSWE}{dt} = a_s - m_s - s_s \quad (1a)$$

$$a_s = \begin{cases} 0, & T \geq T_{accm} \\ p, & T < T_{accm} \end{cases} \quad (1b)$$

$$m_s = \begin{cases} M_f(T - T_{melt}), & T \geq T_{melt} \\ 0, & T < T_{melt} \text{ or } SWE = 0 \end{cases} \quad (1c)$$

$$s_s = 0 \quad (1d)$$

where SWE is snow water equivalent (mm), a_s is the rate of snow accumulation (mm day⁻¹), m_s is the snow melt rate (mm day⁻¹), s_s is the rate of sublimation (mm day⁻¹), p is the precipitation rate (mm day⁻¹), T is air temperature (K), M_f is the melt factor (mm K⁻¹ day⁻¹), T_{accm} is the temperature threshold to distinguish between rain and snow (K), and T_{melt} is the temperature threshold for snow melt (K). In its most parsimonious form $T_{accm} = T_{melt}$, and the model has only two parameters.

There are clearly several simplifying assumptions in this basic snow model. First, the use of a single value to distinguish between rain and snow does not account for mixed-phase precipitation or the transition between rain and snow over a model time step. It is relatively simple to extend the model to simulate a mix of rain and snow over a time step; for example, through the use of a normal distribution, $N(T_{accm}, \sigma_{accm}^2)$, where the fraction of rainfall is equal to the cumulative probability of the temperature T . However, accounting for a mix of rain and snow requires specifying or estimating an additional model parameter (σ_{accm}^2). Second, the assumption of zero sublimation (Eq. 1d) may be unrealistic in some snow environments. The assumption is, however, somewhat justified in New Zealand, because the warm, humid conditions mean that large negative moisture gradients between the snow and atmosphere are uncommon – in New Zealand the latent heat flux is often positive (energy is directed towards the snow surface), resulting in condensation rather than sublimation (e.g., Moore and Owens, 1984). Third, the constant relationship between melt and temperature (Eq. 1c) ignores many important melt processes. We will explore temporal variability in the melt factor in subsequent sections.

SnowSim extensions to the basic snow model

SnowSim includes temporal variability in the melt factor to account for (i) seasonal changes in the energy available for melt that are unrelated to air temperature; (ii) reduction in melt due to the higher albedo of fresh snow; and (iii) enhanced melt during rain-on-snow events. This is parameterized as follows (McAlevey, 1998; Kerr, 2005):

$$M_f = M_f^{\min} + \frac{M_f^{\max} - M_f^{\min}}{1 + \exp(2.5 - 0.2\alpha t)} \quad (2)$$

where M_f^{\min} and M_f^{\max} are the minimum and maximum melt factors (mm K⁻¹ day⁻¹), α is a dimensionless rain-on-snow parameter ($\alpha = 1$ when there is no precipitation and $\alpha = 2$ when precipitation is greater than zero), and t is the time (in days) since the last snowfall. The minimum melt factor is set to 3 and the maximum melt factor M_f^{\max} varies seasonally ($M_f^{\max} = 8$ mm K⁻¹ day⁻¹ for the period 1st April to 5th December, $M_f^{\max} = 9$ mm K⁻¹ day⁻¹ for the period 6th December to 5th January, $M_f^{\max} = 10$ mm K⁻¹ day⁻¹ for the period 6th January to 25th January, and $M_f^{\max} = 12$ mm K⁻¹ day⁻¹ for the period 26th January to 31st March).

Figure 1 illustrates how the melt factor changes with time since the last snowfall, as predicted by the *SnowSim* parameterization (Eq. 2), and exposes two main inconsistencies in the *SnowSim* approach. First, albedo is parameterized using a logistic function (Eq. 2), which decreases snow albedo relatively slowly after fresh snowfall, whereas in nature albedo is known to decrease rapidly after fresh snowfall (e.g., Colbeck, 1982). Second, the enhanced melt during rain-on-snow events is included as part of the albedo parameterization, so that the amount of additional melt simulated during rain-on-snow events depends on the time since the

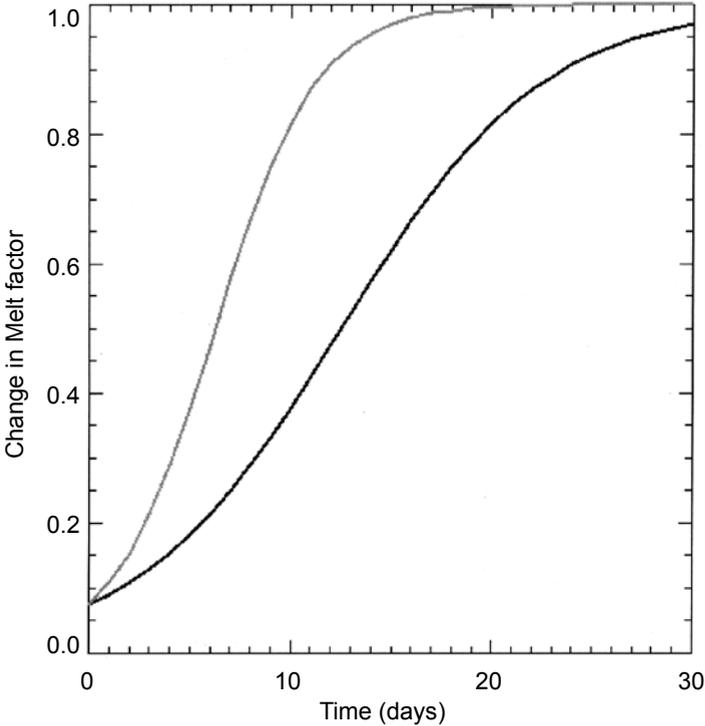


Figure 1 – Change in melt factor with time since the last snowfall, as computed using $1/[1 + \exp(2.5 - 0.2\alpha t)]$ (see Eq. 2). The dark line is for the case $\alpha = 1$ (zero precipitation) and the light line is for the case $\alpha = 2$ (rain-on-snow events).

last snowfall. Therefore, as demonstrated in Figure 1, *SnowSim* has the unrealistic behavior that enhanced melt during rain-on-snow events is zero both immediately after fresh snowfall and when snow is completely “ripe” (e.g., 10 days after a snowfall event). These inconsistencies in *SnowSim* may not lead to poor simulations in all circumstances, but should be resolved to improve our general level of confidence in the model output.

Another aspect that deserves attention is the method used in *SnowSim* to parameterize seasonal variability of the melt factor. *SnowSim* represents this as a series of step changes throughout the snow season (at April 1st, December 6th, January 6th, and January 26th). This is conceptually unsatisfying because seasonal changes in the energy available for melt do not typically occur as a series of step changes; in addition the dates for the step changes seem somewhat arbitrary. The parameterization also has practical limitations, in that eight model parameters

must be specified (four step changes in the melt factor at four different dates). It seems more reasonable to model the seasonal variability of the melt factor as a continuous function, such as the sine curve used by Anderson (1973).

Alternative parameterization of temporal changes in the energy available for melt

The limitations with the *SnowSim* parameterization outlined above suggest the need for using alternative parameters of temporal changes in the energy available for melt. In doing so, we will retain the process complexity of *SnowSim* by accounting for (i) seasonal changes in the energy available for melt that are unrelated to air temperature, (ii) reduction in melt due to the higher albedo of fresh snow, and (iii) enhanced melt during rain-on-snow events. We parameterize the melt factor using the formulation:

$$M_f = \max(\bar{M}_f + \delta M_f^{seas} + \delta M_f^\alpha + \delta M_f^{ros}, 0) \quad (3a)$$

where \overline{M}_f is the mean melt factor, δM_f^{seas} accounts for the change in melt factor due to seasonal variability in the energy available for melt, δM_f^α accounts for the reduction in the energy available for melt due to the higher snow albedo immediately after fresh snow, and δM_f^{ros} accounts for additional melt during rain-on-snow events. Note that the melt factor is constrained to be positive.

Seasonal variability in the melt factor can help to compensate for temporal variability in the energy available for melt that may be missing from the snow model; for example, seasonal changes in solar radiation and the energy used to warm the snow to 0°C. Following Anderson (1973), we parameterize the seasonal variability as a sine curve

$$S_i = \begin{cases} -\sin\left(\frac{d2\pi}{366} + 0.551\pi\right), & lat > 0 \\ \sin\left(\frac{d2\pi}{366} + 0.551\pi\right), & lat < 0 \end{cases} \quad (3b)$$

where S_i is the solar index, d is the number of days since January 1st of the current year, and the offset 0.551π is used to ensure the melt factor is at its lowest on 21 December in the Northern Hemisphere ($lat > 0$) and on 21 June in the Southern Hemisphere ($lat < 0$). The seasonal change in melt factor (δM_f^{seas}) can then be computed as

$$\delta M_f^{seas} = M_f^{amp} S_i \quad (3c)$$

where M_f^{amp} (mm K⁻¹ day⁻¹) is the seasonal amplitude of the melt factor.

The change in the melt factor associated with higher albedo is parameterized as:

$$\delta M_f^\alpha = -M_f^\alpha \exp\left(-\frac{d_{snow}}{r}\right) \quad (3d)$$

where d_{snow} is the time since fresh snowfall (days), r is the timescale for the decrease in snow albedo (days), and

M_f^α (mm K⁻¹ day⁻¹) is the reduction in melt associated with the higher snow albedo immediately after fresh snowfall.

Finally, we parameterize melt during rain-on-snow events simply as:

$$\delta M_f^{ros} = \begin{cases} 0, & p - a_s = 0 \\ M_f^{ros}, & p - a_s > 0 \end{cases} \quad (3e)$$

where M_f^{ros} (mm K⁻¹ day⁻¹) is the additional melt during rain-on-snow events. The melt during warm rain-on-snow events has been observed to be twice as much as the melt during anticyclonic conditions. For example, Moore and Owens (1984) observed that snow melt at Temple Basin (near the Main Divide of the Southern Alps) was ~200 mm/day during strong northwesterly (föhn) rain events, but was only 100 mm/day during anticyclonic conditions. However, melt rates during southwesterly rain events was observed to be similar to melt rates during anticyclonic conditions (Moore and Owens, 1984). An appropriate default value for M_f^{ros} is therefore approximately half of the value assigned to \overline{M}_f , but such a simple parameterization of the energy available for melt does not account for differences in wind speed and heat advection in different synoptic events.

This alternative snow model has seven parameters (Table 1, page 46).

Model configuration and input data

The new snow model detailed in Section 2.3 is configured to run for all 20,040 third-order river basins in the South Island (as delimited in the River Environment Classification; Snelder and Biggs, 2002). When discretized into 100-metre elevation bands, we obtain 258,307 computational elements (see Fig. 2). The model was forced using NIWA's

Table 1 – Snow model parameters

Parameter	Description	Default	Units
T_{accm}	Temperature threshold for snow accumulation	274.16	K
T_{melt}	Temperature threshold for snow melt	273.16	K
\overline{M}_f	Mean melt factor	5	mm K ⁻¹ day ⁻¹
M_f^{amp}	Seasonal amplitude in the melt factor	5	mm K ⁻¹ day ⁻¹
M_f^α	Decrease in melt factor due to higher fresh snow albedo	2.5	mm K ⁻¹ day ⁻¹
r	Timescale for decrease in snow albedo	5	day
M_f^{ros}	Increase in the melt factor in rain-on-snow events	2.5	mm K ⁻¹ day ⁻¹

gridded daily surfaces of precipitation and maximum and minimum temperature for the period 1997–2005. Precipitation surfaces were produced by using a second-order derivative thin plate smoothing spline to interpolate point station data onto a regular 0.05° latitude-longitude grid (approx. 5 km × 5 km) across New Zealand (Tait *et al.*, 2006). Temperature surfaces were produced

by first extrapolating the station data to sea level (lapse rate of 5 K km⁻¹) before applying the thin-plate spline model, and by then extrapolating the interpolated temperature back to the elevation of the 0.05° latitude-longitude grid. The lapse rate of 5 K km⁻¹ is consistent with the estimate of Norton (1985) and reflects the humid conditions in New Zealand.

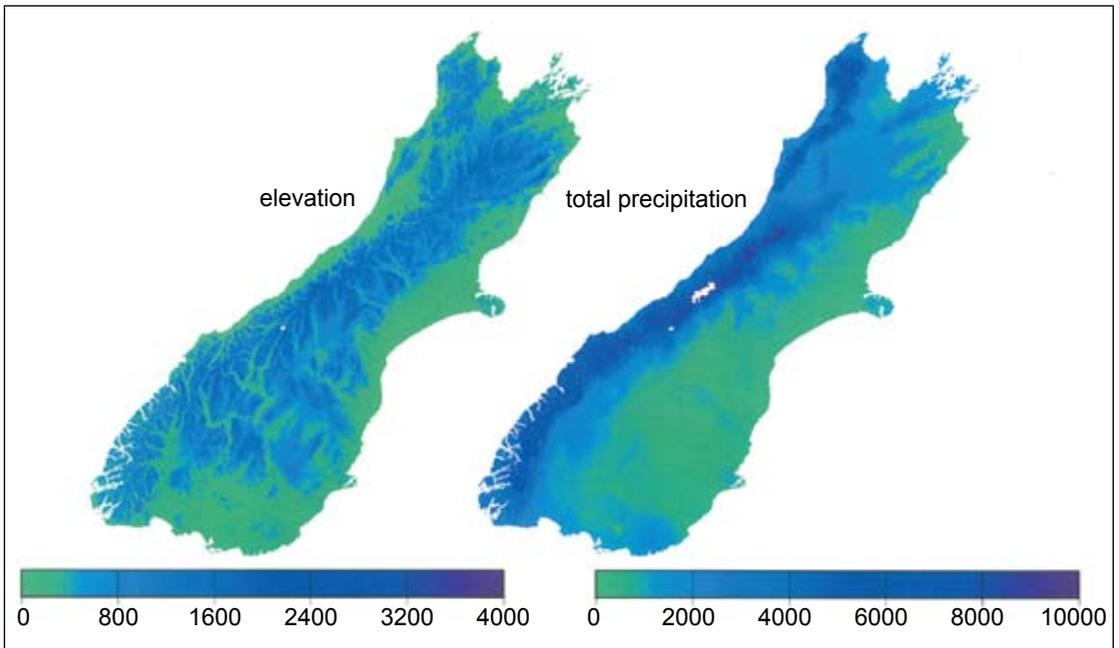


Figure 2 – Elevation (metres above sea level) and total annual precipitation (mm year⁻¹) for each sub-basin in the modelling domain. The sub-basins shaded white have precipitation >10,000 mm year⁻¹.

The snow model uses the interpolated climate data from the closest grid cell to each sub-basin as the model forcing for that sub-basin. This is done because the spatial scale of each sub-basin is approximately equivalent to the horizontal resolution of the gridded climate surfaces (5 km × 5 km), and many of the key processes (e.g., orographic enhancement) occur at spatial scales larger than 5 km. By doing this, the credibility of our model simulations depends critically on the credibility of the gridded climate surfaces – while these climate surfaces are still far from perfect (Tait *et al.*, 2006), they do describe key the features on New Zealand’s climate such as orographic enhancement of precipitation and strong spatial gradients in precipitation from the windward to leeward side of the Main Divide. The gridded precipitation data are used without modification, but the temperature data are adjusted to the mid-point of each elevation band using a lapse rate of 5 K km⁻¹.

All model simulations were performed at an hourly time step, which requires temporal disaggregation of the daily gridded climate estimates. The daily precipitation data were disaggregated to hourly time intervals using the multiplicative random cascade method (Rupp *et al.* 2009), in which rainfall R occurring over an interval in time is divided among a number of smaller intervals of equal size (we use two sub-intervals). The rain that falls into any particular subinterval is determined by multiplying the rain R by a dimensionless random variable W , or splitting weight, and the process is repeated until the desired temporal resolution is reached. Over many ‘cascade’ levels a highly intermittent time series resembling a real rainfall record is generated. Hourly temperature data were produced by fitting a sine curve to the spatially interpolated maximum and minimum temperatures, in which the time of maximum temperature was determined by fitting the parameters of

the sine curve to hourly station data (Clark *et al.*, 2008).

Snow information

Historically, there has been no systematic programme of seasonal snow-monitoring in New Zealand, and so extensive quantitative evaluation of the model simulations is problematic. In this study, we evaluate the snow simulations using three pieces of information:

1. *Measurements of snow water equivalent*
Table 2 (page 48) summarizes data on maximum seasonal snow water equivalent reported in previous studies in New Zealand. Maximum snow water equivalent is often greater than 500 mm, especially at elevations greater than 1600 metres.
2. *Water balance estimates of snow storage*
Fitzharris and Grimmond (1982) and McKerchar *et al.* (1998) estimated snow storage as equivalent to the residual in the monthly water balance, i.e., given $Q = P - E - \Delta S$ (where Q is runoff, P is precipitation, E is actual evaporation, and ΔS is the change in storage), $\Delta S = P - E - Q$. Using this approach, Fitzharris and Grimmond (1982) estimate seasonal snow storage in the Fraser catchment in Central Otago to be 176 mm and McKerchar *et al.* (1998) estimate average seasonal snow storage in the Waitaki catchment above Lakes Tekapo, Pukaki and Ohau to be 500 mm, and in the Clutha catchment above Lakes Hawea, Wanaka, and Wakatipu to be 300 mm. McKerchar *et al.* (1998) also estimate the percentage of annual runoff that is snowmelt to be 22% in the Waitaki catchment and 16% in the Clutha catchment.

Following an approach similar to that mentioned above, estimates of mean snow storage for the period between January 1970 and December 2001 were

Table 2 – Snow water equivalent (SWE) measurements in New Zealand

Reference	Region	Elevation	Year	SWE (mm)	
Morris and O’Loughlin, 1965 O’Loughlin, 1969	Upper Broken River, Craigieburn Range	1432	1962	100	
			1963	280	
			1964	250	
			1965	270	
			1966	130	
			1967	130	
			1968	380	
			1962	560	
			1963	660	
			1964	710	
Fitzharris, 1979	Tasman glacier	1707	1965	630	
			1966	400	
			1967	200	
			1968	1010	
			1550	1100	
			1750	2300	
Owens, 1979 ^a	Rastusburn, Remarkables Range	1800	1976	1825	3100
				1975	4100
				2050	4800
				2350	5800
				1979	0
Weir, 1979	Mount Hutt	1500	1979	1420	459
				1500	484
				1680	779
				1790	916
				1940	763
				2040	303
				Harrison, 1986	Fraser River, Central Otago
1976	310				
Fitzharris and Garr, 1995	Mueller Hut, Mount Cook	1600	1993	1200	
	Mt. Belle, Fiordland	1650	1993	1300	

^a Data presented in Harrison (1986).

calculated for 468 catchments across the South Island using mean monthly values of precipitation and potential evapotranspiration (PET) calculated from daily spline interpolations generated by Tait *et al.* (2006) and Tait and Woods (2007), and mean runoff from Woods

et al. (2006). Our method for calculating ΔS explicitly diagnoses and corrects inconsistencies between precipitation and runoff data (computational details are provided in the Appendix). Figure 3 illustrates water balance estimates of the sum of positive change in storage (after

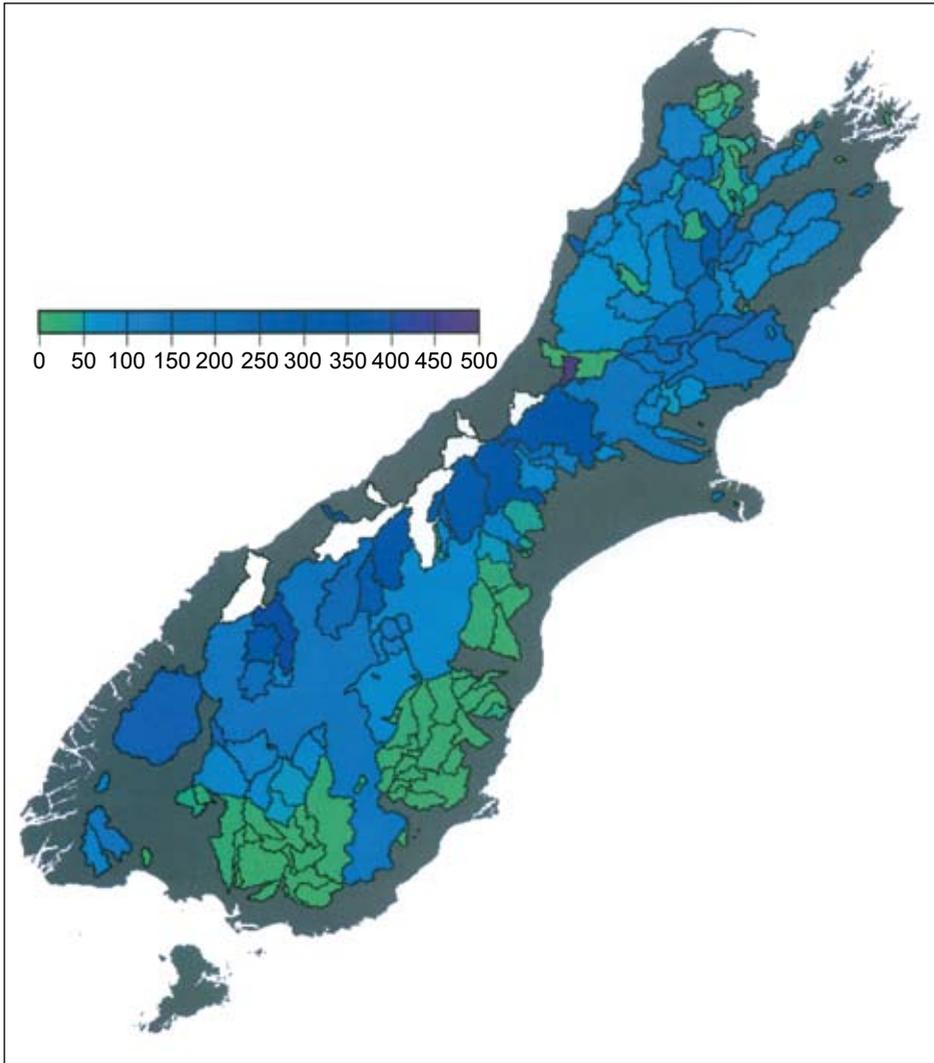


Figure 3 – Water-balance estimates of total snow storage (mm) at the end of September for major South Island catchments (i.e., total cumulative storage minus estimates of plant available water in the soil, as provided by the New Zealand Land Resource Inventory). The end of September is assumed to be the month representing the maximum snow storage. Grey areas are where there are no streamflow measurements, and white areas are where the catchment-average storage is greater than 500 mm.

removing plant-available storage in the soil), for catchments in the South Island in the month of September.

3. *Classification of snow regions* Figure 4 shows the snow regions in New Zealand, as classified by the Technical Subcommittee

on Snow (1969). They define seasonal snow as ‘where snow persists for an appreciable period in the spring, say to September in most years.’ Seasonal snow in the Technical Subcommittee classification covers 35% of the land area of the South Island (Owens and Fitzharris, 2004). For the

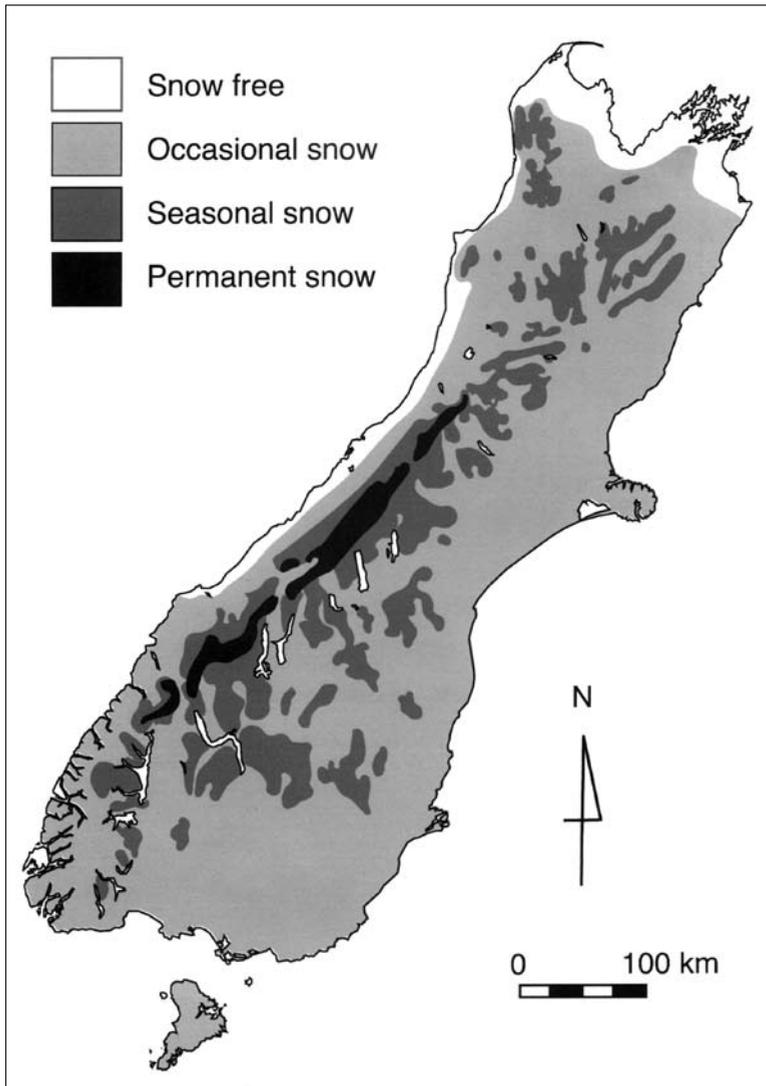


Figure 4 – Classification of snow by the Technical Subcommittee on Snow (1969).

quantitative comparisons needed for this paper, the duration of seasonal snow is assumed to be at least three months (land is snow-covered for 25% of the year).

Restricting attention to these three data sources means that the model evaluation is limited to assessment of the model’s snow climatology. Another source of data available for model evaluation is satellite-based

estimates of snow-covered area. However, these satellite-based estimates are of limited use for validating large-scale models, because (i) the standard snow products have severe problems discriminating between snow and clouds(e.g., Sirguey *et al.*, 2009); and (ii) visible satellite imagery does not provide any information when the snow is obscured by clouds, and this is often the time when

melt rates are highest (e.g., during rain-on-snow events; Moore and Owens, 1984). While specialized snow products have been developed that address problems of snow-cloud discrimination (e.g., Sirguy *et al.*, 2009), these products have a limited spatial domain (i.e., individual river basins) and hence cannot be used to evaluate snow simulations for the South Island as a whole. Detailed comparison of model output to satellite-based estimates of snow-covered area at the scale of individual basins will be pursued in future work.

Model simulation results

Figures 5-7 illustrate, respectively, the sensitivity of the snow model parameters to (i) the fraction of precipitation that falls as snow (Fig. 5); (ii) snow duration (Fig. 6); and (iii) the maximum snow mass in each year, averaged over the 9-year simulation period (Fig. 7). Maps are shown for T_{accm} parameters ranging from -1°C to $+2^{\circ}\text{C}$,

and for melt parameters of $(\bar{M}_f, M_f^{amp}) = (5 \text{ mm K}^{-1} \text{ day}^{-1}, 5 \text{ mm K}^{-1} \text{ day}^{-1})$, $(2.5 \text{ mm K}^{-1} \text{ day}^{-1}, 2.5 \text{ mm K}^{-1} \text{ day}^{-1})$, and $(5 \text{ mm K}^{-1} \text{ day}^{-1}, 2.5 \text{ mm K}^{-1} \text{ day}^{-1})$. In all simulations, the parameters T_{melt} , M_f^{α} , r , and M_f^{pos} are set to their default values (Table 1). Figure 5 includes only one set of melt parameters because the melt parameters have no impact on the fraction of snowfall. The snow duration maps (Fig. 6) can be compared directly with the snow classification presented in Figure 4, and the maximum snow mass (Fig. 7) is presented as a proxy for snow storage, and can be compared against the measurements of snow storage in Table 2 and the water balance estimates of snow storage in Figure 3.

The statistics presented in Figures 5-7 are broadly consistent with observations: the largest simulated snow volumes are along the main Divide of the Southern Alps (Fig. 7), and seasonal snow is present throughout mountainous regions (Fig. 6). However,

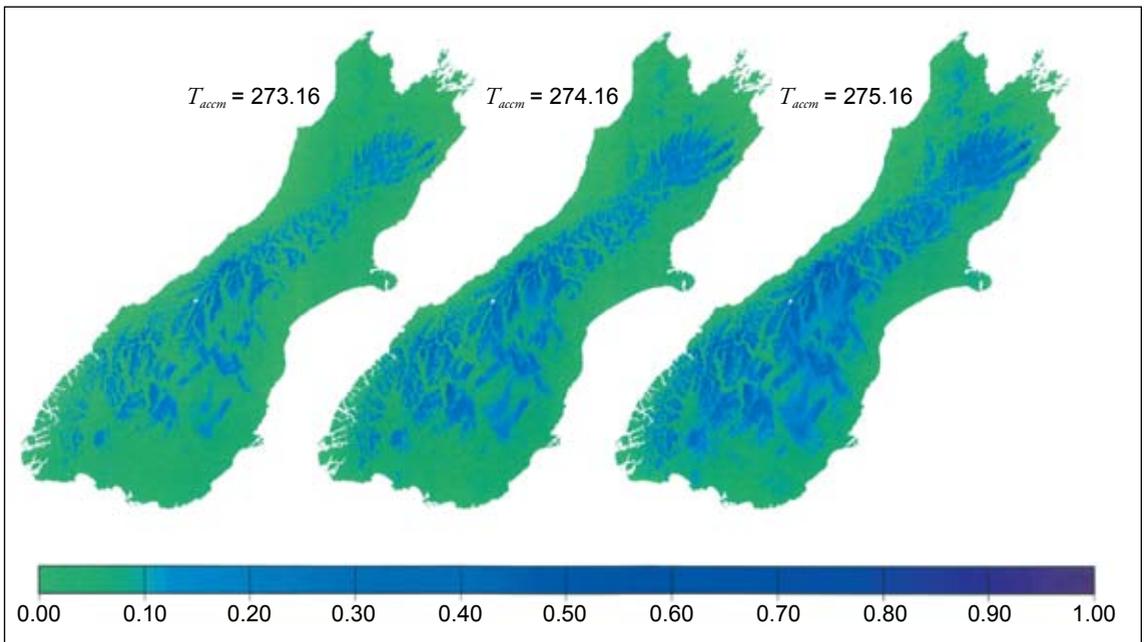


Figure 5 – Fraction of total precipitation that is snowfall for different T_{accm} parameters. Note that the melt parameters $(\bar{M}_f$ and $M_f^{amp})$ have no impact on the fraction of snowfall.

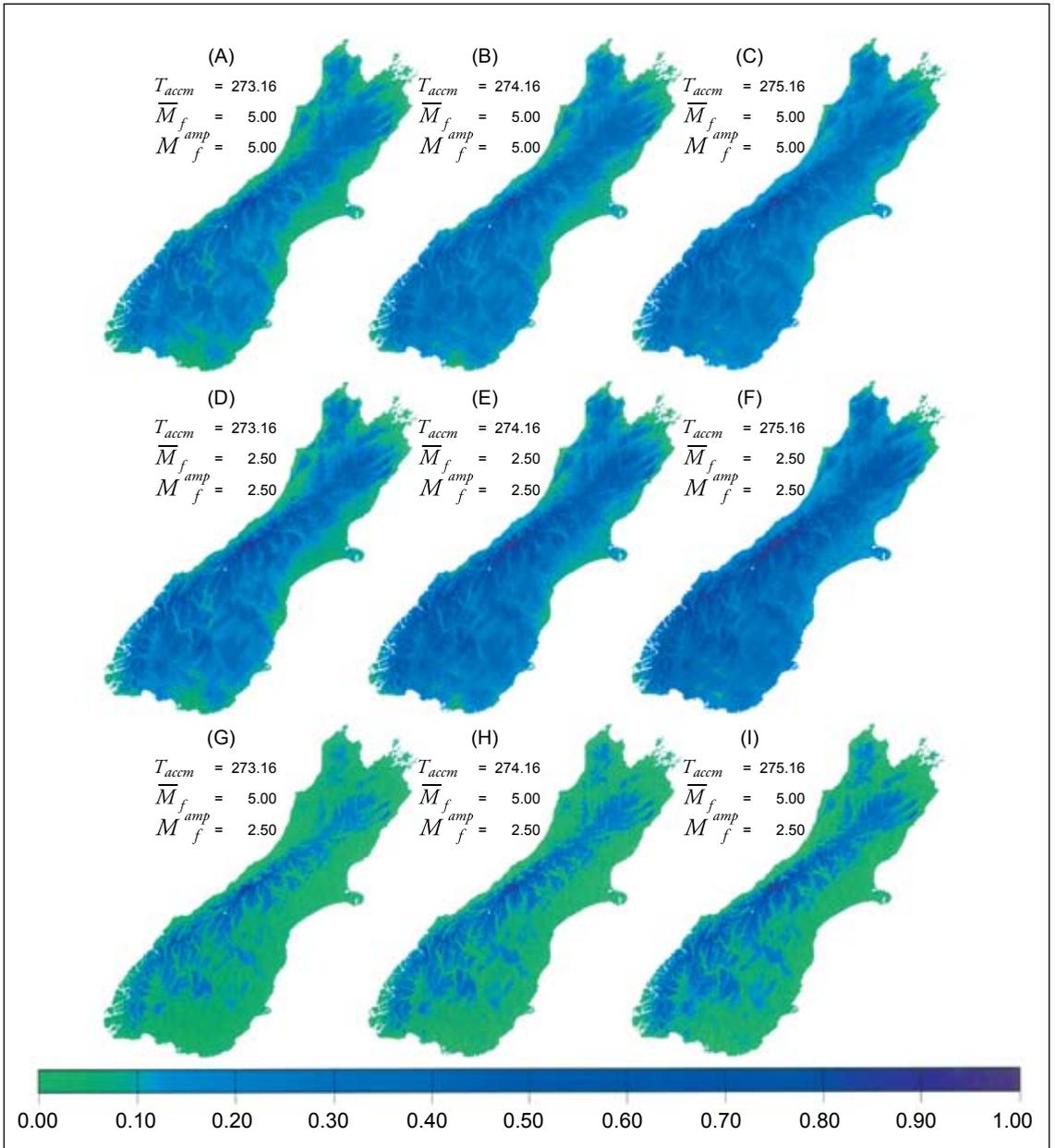


Figure 6 – Fractional snow duration for different model parameter sets. As suggested by qualitative comparison of the mapping of seasonal snow in Figure 4, plots C, E, and F have excessive snow duration and plots G, H, and I have insufficient snow duration.

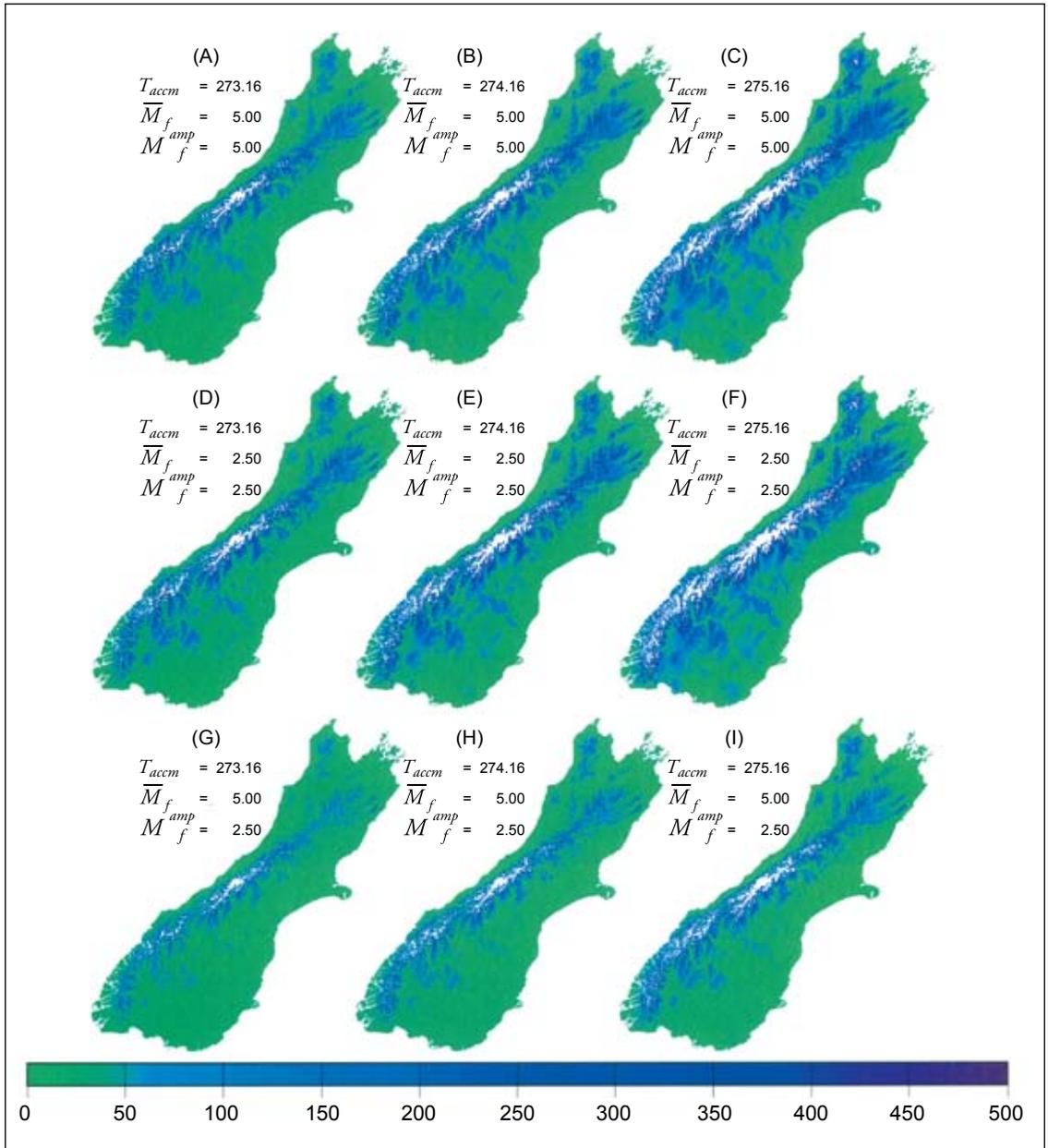


Figure 7 – Average maximum snow accumulation (mm) for different model parameter sets. Plots A, D, G, and H have insufficient snow volume, as suggested by qualitative comparison of the water balance estimates of snow storage in Figure 3.

there are striking behavioral differences between the various parameter sets tested, which should allow us to eliminate some of the model parameter sets based on the snow information presented in Figures 3 and 4 and in Table 2.

The parameter sets with $\overline{M}_f = 5 \text{ mm K}^{-1} \text{ day}^{-1}$ and $M_f^{amp} = 2.5 \text{ mm K}^{-1} \text{ day}^{-1}$ cause a substantial fraction of the accumulated snow to melt in the middle of winter. Hence the duration of snow is much lower than the Technical Subcommittee on Snow classification (compare the bottom row of Figure 6 with Figure 4) and the volume of snow is lower than the water-balance estimates (compare the bottom row of Figure 7 with Figure 3). With parameter sets $\overline{M}_f = 5 \text{ mm K}^{-1} \text{ day}^{-1}$ and $M_f^{amp} = 2.5 \text{ mm K}^{-1} \text{ day}^{-1}$, modelled snow storage greater than 500 mm is restricted to a thin ribbon along the main divide of the Southern Alps (Fig. 7), whereas the water balance estimates of snow storage indicate snow volumes in excess of 500 mm occur over large catchments (Fig. 3).

The parameter sets where $T_{accm} = 2^\circ\text{C}$ result in a large fraction of precipitation falling as snow (right plot in Figure 5) and a duration of snow that is much higher than is shown by the Technical Subcommittee on Snow classification (compare the right column of Figure 6 with Figure 4). Conversely, the parameter sets where $T_{accm} = 0^\circ\text{C}$ result in a much smaller fraction of precipitation falling as snow (left plot in Figure 5) and insufficient snow volume (compare the left column of Figure 7 with Figure 3).

Taken together, the maps in Figures 5–7 suggest appropriate default values for the primary snow model parameters are somewhere in the vicinity of $T_{accm} = 1^\circ\text{C}$ and $\overline{M}_f = M_f^{amp}$ (Table 1). The equality

$\overline{M}_f = M_f^{amp}$ causes melt to approach zero in the middle of winter.

We conducted additional simulations (not shown) to assess the sensitivity of the snow model to other snow model parameters (T_{melt} , M_f^α , r , and M_f^{ros}). The impact of those other snow model parameters is difficult to distinguish from the impact of T_{accm} and \overline{M}_f and M_f^{amp} . For example, the reduction in melt associated with albedo (determined by parameters M_f^α and r) depends on the frequency of snowfall (determined by parameter T_{accm}), and parameters M_f^α and r can hence have a similar impact on the seasonality in melt to the parameter M_f^{amp} . Also, the impact of T_{melt} is not clearly separable from the impact of \overline{M}_f . These parameter interactions hamper efforts to identify unique parameter values.

Summary and discussion

This paper presents snow simulations for the South Island, New Zealand, using different model parameter sets. These simulations appear to provide an adequate representation of New Zealand's snow climatology. While the use of more physically-based energy-balance methods may help to improve the simulations (Slater *et al.*, 2001; Etchevers *et al.*, 2004), such methods have larger data requirements, and it is unclear whether the improvements so obtained would be significant given the uncertainty in the model input data (especially when the input data is interpolated over large horizontal distances, i.e., from stations to sub-basins). Further research is needed to evaluate different melt parameterizations.

The accuracy of snow simulations can be improved by optimally combining model output with observations. This can be done by assimilating point snow water equivalent observations and satellite-derived estimates of snow-covered area into the model (Fitzharris

and McAlevey, 1999; Clark *et al.*, 2006; Slater and Clark, 2006). Data assimilation will provide snow information that takes advantage of the relative strengths of data and models, and the merged data-model product will be of greater accuracy than either the data or model alone. However, data assimilation will only improve present-day simulations, and there is a pressing need to identify appropriate model equations and model parameter sets that can be used for future scenarios, for example, to simulate the impact of climate change on water resources.

The problem of model structure identification and model parameter estimation is clearly still unresolved. Lack of data for model evaluation is a ubiquitous problem in hydrology, and there is increasing recognition of the need to compare model output to multiple quantitative and qualitative data sources (e.g., Gupta *et al.*, 1998). Indeed, there is an emerging research trend to creatively use all available data in such a way that the data provides information on both the behavior of model sub-components and appropriate values for individual model parameters (e.g., Gupta *et al.*, 2008). For example, qualitative information on snow duration across the South Island (as assessed

in this paper) can provide more information on appropriate values of melt parameters than quantitative data on snow water equivalent at a single station location. Further efforts in improving snow models in New Zealand, that is, diagnosing and correcting model weaknesses or regionalizing model parameter sets, will require creative assessment of a much broader set of information than the statistics on seasonal snow climatology considered in this paper. Important data sources for future studies will be both satellite data on snow-covered area (e.g., as produced by Sirguey *et al.*, 2009, as preliminary analysis indicates that standard satellite snow products have problems with snow-cloud discrimination) and seasonal and diurnal cycles in streamflow. Future research on improving snow models will improve predictions on water availability and guidance on likely changes in New Zealand's snow resource.

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Appendix: Water balance estimates of snow storage

Following Fitzharris and Grimmond (1982) and McKerchar *et al.* (1998), we use a water balance method to provide alternative estimates of snow storage.

The change in storage can be calculated from long-term average monthly data as

$$\Delta S = P - E - Q \quad (\text{A.1})$$

where Q is runoff, P is precipitation, E is actual evaporation, and ΔS is the change in storage. Note that if the basin-average precipitation estimates are biased, as may occur because of poor spatial representativeness of the available stations as well as

possible data errors at individual stations, the estimates of ΔS are questionable.

Potential problems with the basin-average precipitation estimates can be diagnosed and corrected using the 'Budyko curve' (Budyko, 1974), which describes the relationship between an *annual* index of dryness, expressed as PET/P , and the partitioning of precipitation between evaporation and runoff, expressed as E/P (where $E = P - Q$). Figure A1 illustrates the relationship between the dryness index and precipitation partitioning for all major catchments in the South Island. Milly and Dunne (2002)

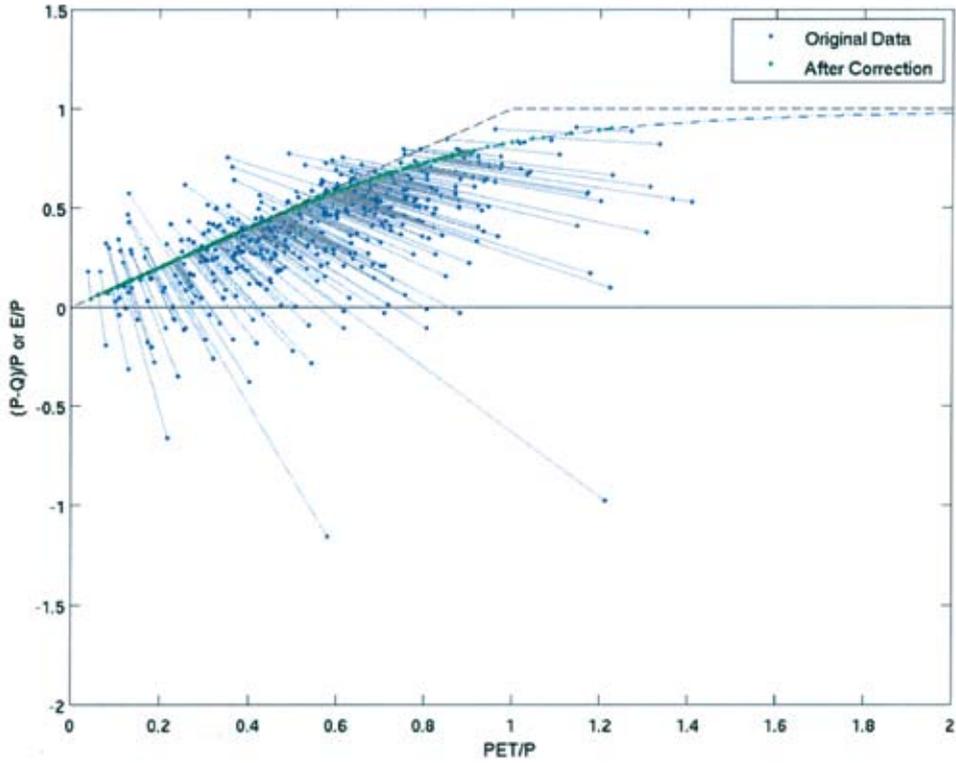


Figure A1 – Relationship between PET/P and E/P for all major South Island catchments.

have demonstrated that ‘poor’ basin-average precipitation estimates have a much larger scatter around the Budyko curve than ‘good’ basin-average precipitation estimates. Consequently, problems with the basin-average precipitation estimates can be diagnosed and corrected if a relationship between PET/P and E/P is assumed.

To this end, we use the Zhang *et al.* (2004) equation to provide a functional relationship between PET/P and E/P ,

$$E_z = PET \left(1 + \frac{P}{PET} - \left[1 + \left(\frac{P}{PET} \right)^w \right]^{1/w} \right) \quad (A.2)$$

where w is an adjustable parameter. This is the same functional form used by Woods *et al.* (2006), and, following Woods *et al.* (2006), $w = 4.35$. The inconsistencies between P and

Q were then resolved by iteratively adding linear adjustments to P until E_z computed by equation matches $E = P - Q$. The linear *annual* precipitation correction factors are then applied to each month.

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