# Chapter 5
# Seasonal snow and water

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Of the total global precipitation, only about 6% falls as snow, so at first glance snow does not seem an important part of hydrology. However, storage of water as snow exerts a powerful influence on runoff at many time scales. Seasonal snow accumulates and then melts within a single hydrological year. In so doing it tends to reduce river flows in winter, but enhances them in spring and early summer.

## SNOW IN NEW ZEALAND

New Zealand lies between latitudes 34° and 47°S in the southwest Pacific. Surrounded by ocean, its main axial ranges trend across the prevailing westerlies, and reach elevations above 2000 m in many places. Most seasonal snow is concentrated in the mountains of the Southern Alps, but it also accumulates on the volcanic cones of the North Island and on its main axial ranges. The combination of high relief and moist on-shore winds ensures that up to 35% (53,000 km$^2$) of the South Island is snow-covered throughout the winter (Fig. 5.1). The winter elevation of the snow line is highly variable, but averages 1000 m in the south and 1400 m in the north. High summer solar radiation levels at New Zealand's mid-latitude location ensures strong melting and a supply of snow water to rivers.

Seasonal snow accumulation exceeds 4000 mm water equivalent near the Main Divide of the Southern Alps, but is generally less than 1000 mm in the eastern mountains (Chinn 1969; Fitzharris 1979). By the end of summer, the snow line retreats to elevations of between 1500 m and 2200 m, which marks the perennial snow margin. Earlier reviews of seasonal snow in New Zealand can be found in Fitzharris *et al.* (1992) and Fitzharris *et al.* (1999).

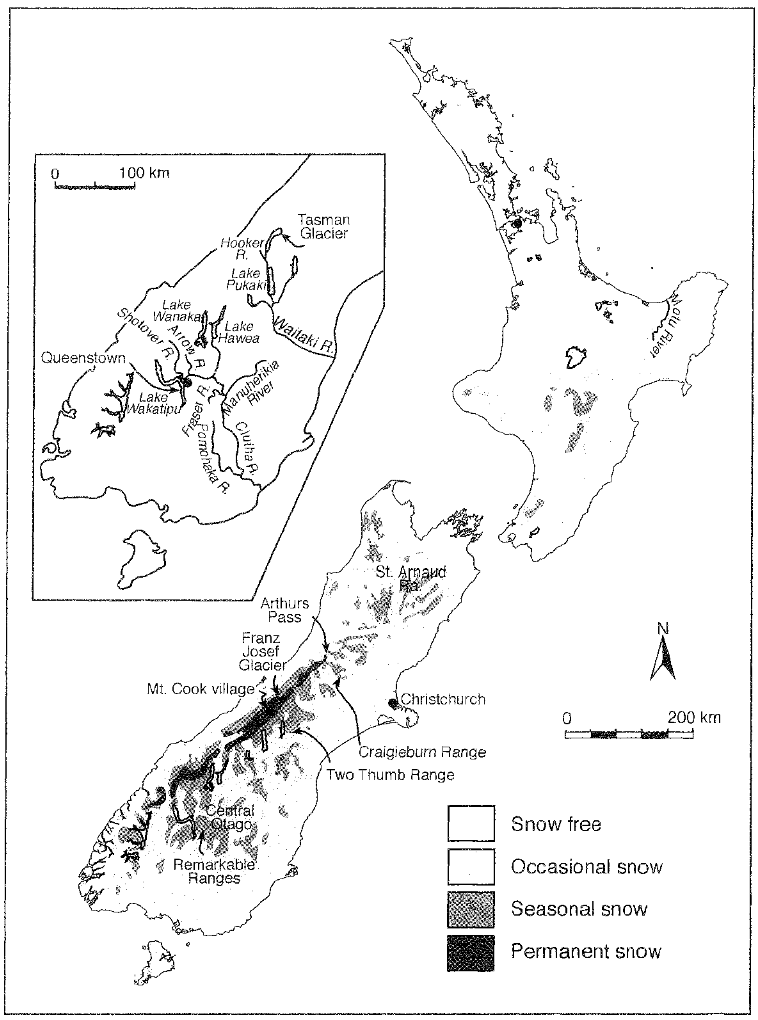


**Figure 5.1** Snow- and ice-covered areas of New Zealand and names of locations referred to in the text (modified from Technical Subcommittee on Snow, New Zealand Committee for the I.H.D. 1969).

# MEASUREMENT OF SNOW

## Snow depth and water equivalent

Many different techniques are used to measure the amount of snow (Fig. 5.2). The most straightforward is to measure snow depth at a point. Accumulated snow depth is assessed with reference to a graduated reference stake, while snow boards allow a measure of new snow fall amounts, as they are usually cleared of snow daily. Snow depth can also be remotely measured, for example with an ultrasonic sensor. However, the enormous spatial variability of snow cover, its transient nature, and great depth in the western mountains, makes comprehensive measurement over a large catchment rather difficult (Chinn 1969).

For hydrological purposes, measurement of snow water equivalent, rather than snow depth, is more desirable and can be obtained in various ways. If a sample of known volume is weighed, the density can be determined (Fig. 5.2a), and the water equivalent given as the product of depth and density. A standard procedure in snow hydrology is to measure water equivalent by sampling the entire thickness of the snow pack with a coring device, such as the Federal Sampler (Fig. 5.2d). Often a snow course is set up, where a series of such samples are made in an area of similar slope, elevation and exposure.

In some cases it is necessary to obtain detailed information about individual layers within the snow pack. In these areas, a snow pit is dug (Fig. 5.3), if possible down to ground level. The observer identifies layers, and measures their thickness, temperature and density. A full snow pit analysis will often involve other parameters such as snow type, hardness and wetness.

## Snow deposition

Snow deposition is obtained from recording precipitation gauges, though a turbulence-damping device such as an Alter shield (Fig. 5.2e) is required in windy environments typical of alpine catchments so as to ensure an accurate catch. A measure of both accumulation and melt is obtained with a snow pillow (Fig. 5.2f) in which the pressure of the snow pack on a reservoir of non-freezing liquid is recorded.

## Area of snow

The areal distribution of snow cover is often synthesised from a large number of point observations, as Chinn (1981) and Hughes (1974) have done for storms affecting low elevations in the eastern South Island. Greater detail of snow distribution is sometimes available when snowstorms affect urban areas, and press reports may also allow the recurrence interval of snow falls of different depths to be estimated (Fig. 5.4).

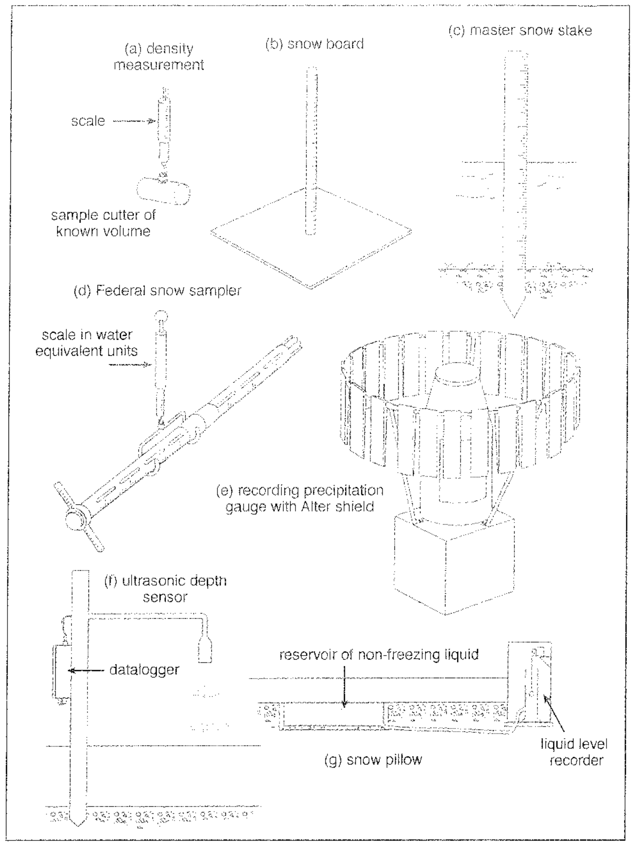


**Figure 5.2** Instruments for measuring snow accumulation and melt.

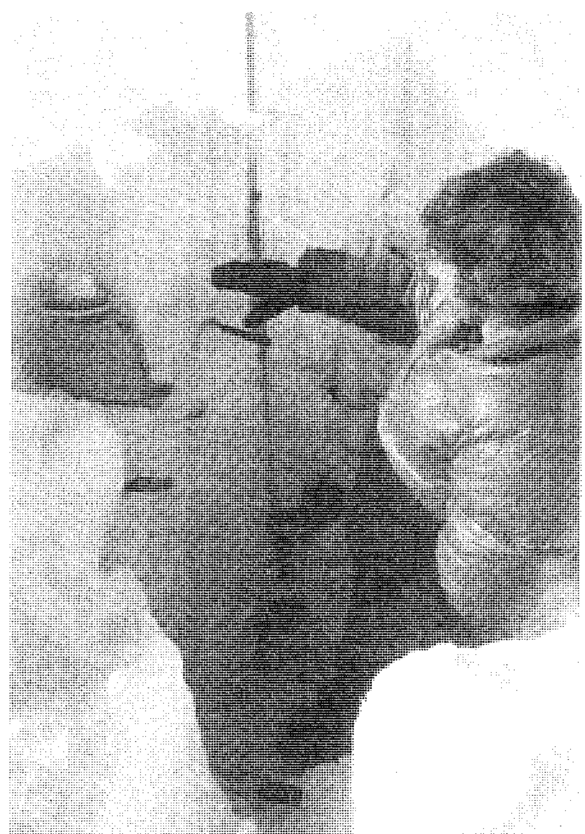

**Figure 5.3** Making observations of snow structure and characteristics within a snow pit.
Photo: Ian Owens

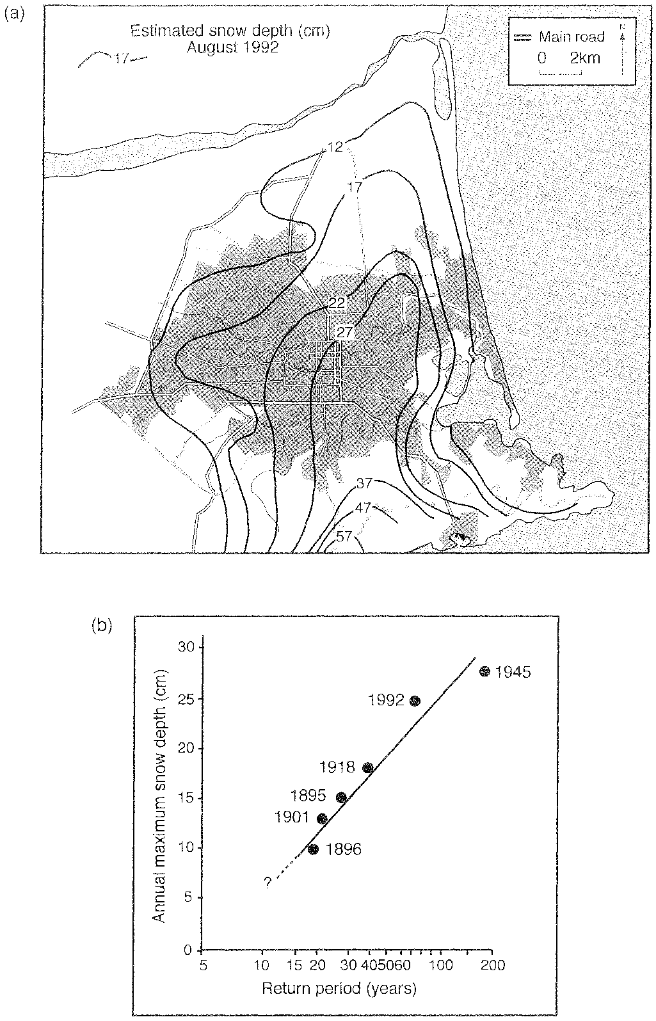


**Figure 5.4** Snow falls in Christchurch. (a) Distribution of snow depths for August 1992 storm (b) Recurrence intervals of snow depth in snow storms for Christchurch (modified from Christchurch Engineering Lifelines Group 1997).

The extent of snow can be determined remotely by aerial photography, or more commonly by using radiation sensors on satellites. An example showing the distribution of seasonal snow over the South Island after a mid-winter southerly blast is given in Figure 5.5. Thomas *et al.* (1978) used Landsat imagery for snow investigations on the St Arnaud Range. Although these images have good resolution (about 15 m), they found them to be of limited use for hydrological applications, because the satellites orbit over the country at 18-day intervals and cloud cover frequently obscures the ground surface. Weather satellites have a much lower resolution (about 1 km), but imagery is available at least daily. The potential for their use was examined by Hickman (1972) and Fitzharris and McGann (1989). They showed that there is reasonable agreement between snow line estimates from satellite imagery and

**Figure 5.5** MODIS satellite image of South Island showing snow cover during the coldest part of the winter (NASA image for 15 June 2002).

from weather stations. Fitzharris and McAlevey (1999) presented three case studies that explore the effectiveness of using weather satellites for determining snow-covered area and snow lines. They concluded that, by itself, satellite monitoring is not sufficient to provide a complete snow climatology. It can, however, give valuable and low-cost monthly checks as to changes in snow cover during the year. They illustrate how, in combination with computer snow simulation models, the area and volume of seasonal snow cover can be obtained.

### Snow melt

Snowmelt can be assessed by repeated measurements of snow depth—the water equivalent is then obtained by multiplying the depth by a standard or known snow density. Alternatively, a lysimeter, placed within or at the base of the snow, catches melt water and measures it through repeated weighing or with a tipping-bucket mechanism at an outlet (Neale and Fitzharris 1997).

## MODELING THE SNOW COVER

If direct measurements of snow depth and water equivalent are not available, snow accumulation is sometimes estimated from standard climate observations. A general index of snow pack accumulation may be derived using daily or monthly deviations from long-term average rainfall and temperatures measured at surrounding lowland climate stations (Fitzharris 1987). Perhaps more effectively, computer simulations can be made of processes of snow accumulation and melt. In this way Moore and Owens (1984) modelled snow water equivalent for the Alan's Basin snow course at 1750 m elevation in the Craigieburn Range, using daily precipitation and temperature data from climate stations located in the nearby mountains. The model requires the optimization of parameters for the rain/snow boundary, a precipitation correction factor to allow for increases with elevation and a degree-day factor for snow melt.

More elaborate definitions of the rain/snow threshold and more detailed specification of temperature variation with height can improve such models. For example, Barringer (1989) simulated snowline elevation on the north-facing slope of the Remarkables Range. He used data from the meteorological station at Queenstown airport and three automatic recording stations at 950 m, 1295 m and 1615 m elevations. Fitzharris and Garr (1995) developed a similar model called *SnowSim*, but which estimates seasonal snow in the main hydroelectric catchments of the South Island. It uses daily climate data from surrounding long-term climate stations. The model is calibrated against the long-term water balance and the output agrees with available historical data.

These simulation models are lumped-parameter models. Apart from variations with elevation, they give no indication of the spatial variability of water stored as snow. McAlevey (1998) extended *SnowSim* to meet this need. His model provides spatial estimates of snow for every 1 $km^2$ pixel for the entire seasonal snow zone of New Zealand. Data is interpolated from 41 climate stations using neural networks, an inverse distance-weighted algorithm and lapse rates for temperature and precipitation. Satellite-derived snow-covered area agrees well with that determined from the model (Fitzharris and McAlevey 1999).

# PHYSICAL PROPERTIES OF SNOW

## New snow types

Most snow originates in the atmosphere at altitudes of 2,000 to 12,000 m, where cloud temperatures are normally below freezing and ice and supercooled water coexist. Ice crystals grow large enough to fall, either by crystal growth (at the expense of water droplets), or from riming (when droplets fall onto crystals and freeze). The exact form of an ice crystal depends on the temperature and humidity conditions at which it grows. The few reported observations for New Zealand mountains, for example by O'Loughlin (1969), Prowse (1981), and Weir and Owens (1981), show that crystals such as heavily rimed needles are very common. Typically, they are associated with humid conditions near the freezing level.

## Cold content of snow

After snow has been deposited, its role in subsequent stages of the hydrological cycle is dependent on the cold content, or heat deficit of the snow pack, which is a function of snow temperature, density and depth. Heat deficits develop in early winter and are not lost until the snow pack becomes isothermal at 0°C in spring. It is then "ripe" for melt. The New Zealand snow pack, however, is relatively warm and may become isothermal at any time. For example, Prowse (1981) calculates that the maximum heat deficit of the Craigieburn Range snow pack could easily be overcome by one day's energy gain, even in winter.

## Snow metamorphism

Before the snow pack becomes ripe, it undergoes metamorphism, which affect its hydrological behaviour (Colbeck *et al.* 1990). If the snow temperature remains lower than 0°C, the snow is said to be dry and metamorphism is controlled by the temperature gradient. Under small gradients, "equilibrium growth" produces rounded, well-bonded snow grains. However, when the temperature change with depth is large, even for small vertical distances, faceted crystals, and eventually hollow cup-shaped crystals called depth hoar, are formed. This process, called "kinetic growth", is common in cold inland locations with shallow snow packs, whereas equilibrium growth is more common in maritime areas with deep, warmer snow packs. Metamorphism of wet snow occurs very rapidly and produces clusters of grains, which eventually loose their identity to form large snow grains known to skiers as spring corn. Refreezing of layers following melting may form very strong and thick ice crusts.

In dry, cold snow, equilibrium metamorphism dominates. Several studies show that depth hoar is surprisingly common in New Zealand, despite our maritime climate. It is observed in colder winters, especially on ranges east of the Main Divide, where snow packs are shallower and cold clear conditions may persist for long periods (McNulty and Fitzharris 1980; Weir and Owens 1981; Prowse and Owens 1984; McGregor 1990). However, most of our snow can be regarded as wet, being at or close to 0°C for much of the time, so that equilibrium metamorphism dominates. Weather processes often create crusts at the snow surface, some of which are subsequently buried.

Apart from indirect effects through avalanche activity, the hydrological significance of snow metamorphism is through the potential of the snow pack to retard delivery of heavy rainfall or large snow melt quantities to river systems. For instance, Moore (1984) showed that liquid water in snow at Temple Basin near Arthur's Pass is trapped and diverted horizontally two to three metres along ice crusts. Nevertheless, measured percolation rates are moderately high because ice layers decay rapidly during rain-on-snow events (Moore and Prowse 1988).

### Snow density

Snow density is the most useful single property for classifying different types of snow and ice phenomena. New snow densities depend on conditions of snow formation and fall. The bulk density of the seasonal snow pack is controlled by processes of metamorphism. In the longer term, firnification leads to eventual formation of glacier ice. Typical densities are given in Table 5.1. At low elevations, new snow densities of about 300 kg/m$^3$ are common (Chinn 1981). At high elevations, most new snow is less dense, but still classified as "damp". Where windy conditions occur, dense slabs are produced from new snow falls (Prowse and Owens 1984).

**Table 5.1** Snow types and their density

| Type | Density (kg m$^{-3}$) |
|---|---|
| Wild snow | 10 – 30 |
| New snow immediately after falling in calm air | 50 – 70 |
| Damp new snow | 100 – 200 |
| Settled snow | 200 – 300 |
| Wind-packed snow - soft slab | 100 – 290 |
| Wind-packed snow - hard slab | 290 – 450 |
| Firn | 400 – 800 |
| Glacier ice | 917 |
| Water | 1000 |

Source: after Seligman(1936), Paterson (1969), Male and Gray (1981)

## PROCESSES OF SNOW ACCUMULATION

The maritime location of New Zealand means that air masses approaching the Southern Alps have freezing levels that are well above sea level. As a consequence, snow lines from storms usually intercept the mountains at some elevation above their base. At the same time, proximity to the cold air of Antarctica ensures that strong southerly flows bring snow to the mountain ranges, even in summer.

With the passage of successive frontal systems, freezing levels frequently rise and fall over many hundreds of metres within a few days, so that snow conditions change rapidly. Precipitation may fall as rain and substantial melt may take place at elevations up to 2500 m or more in mid-winter, yet snow can fall to below 500 m in summer. Thus snow accumulation is not always a steady process, and snow fall or melt occur in every season. The typical evolution of a snow pack is illustrated in Figure 5.6, which shows detailed measurements over a snow season made at two pit sites on the Round Hill Ski Field, Two Thumb Range, Tekapo. Snow falls are related to storm precipitation, as measured a few kilometers away at Mount John, and to the elevation of the freezing level. Each snow fall makes a step increment to the snow-pack depth. Individual snow layers thin exponentially with compaction and metamorphism. Towards October, when the freezing level rises, few additions are made to the snow pack, as precipitation mainly falls as rain.

Processes of deposition and melt create a pattern of increase of snow accumulation with elevation above the snow line that usually resembles a wedge (Fig. 5.7). The shape and slope of the snow wedge are seldom consistent from winter to winter, and are dependent on the frequency of weather types. As snow crystals fall they are deposited on mountains. Those that pass below the freezing level begin to melt, finally turning into rain. Some snow in the layer immediately below the freezing level does not melt

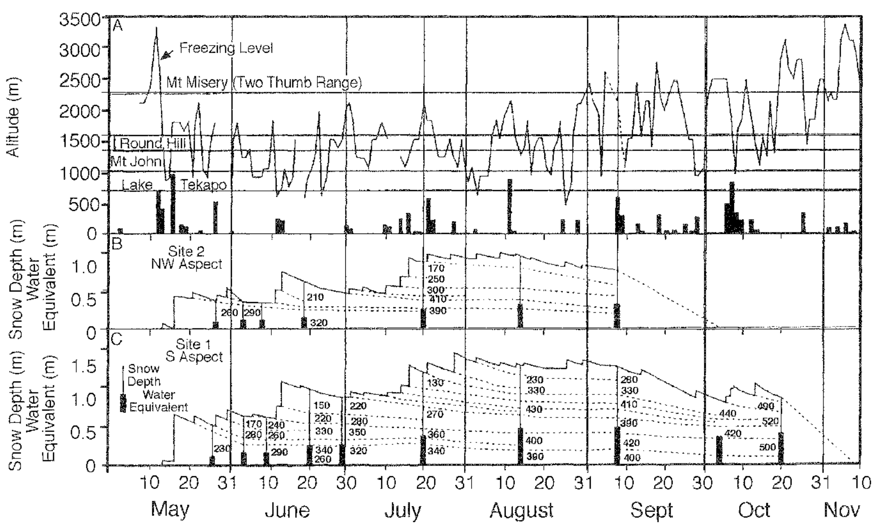


**Figure 5.6** Detailed growth and decline of a snow pack for a typical eastern mountain range: Round Hill, Tekapo, snow course. (Fitzharris *et al.* 1992)

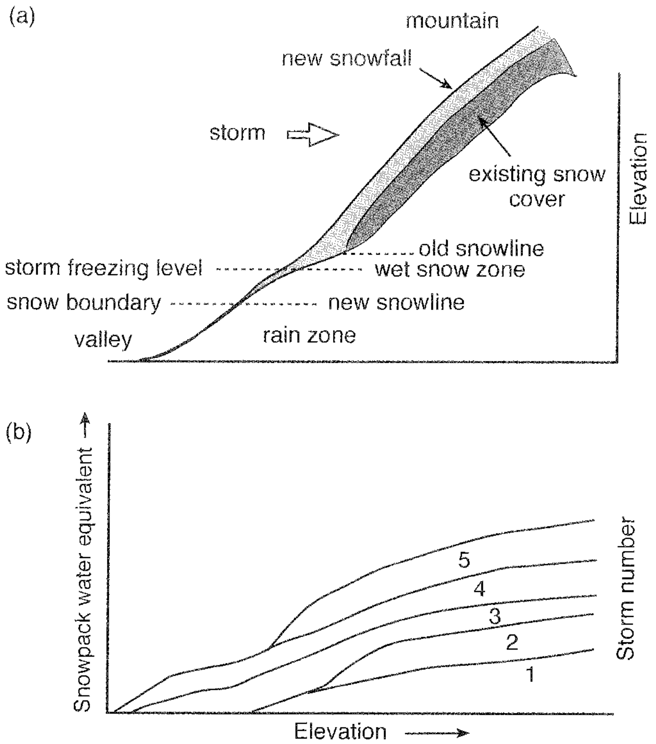


**Figure 5.7** Development of the seasonal snow wedge.

completely before it reaches the ground, and so forms a wet snow zone. The net result is that after a typical storm there is no snow in mountain valleys, but there exists a well-defined fresh snow line part way up the mountain slope. New snow depths increase steadily with elevation in the shape of a snow wedge (Fig. 5.7a).

Along the Southern Alps a typical frontal system creates a compound snow wedge—a higher elevation one is produced by snow in the pre-frontal airstream, which tends to have a higher freezing level; superimposed on this is a second wedge with a lower elevation snow line formed by colder air behind the front. Each storm in the normal sequence of westerly circulation patterns creates its own snow wedges of different shape. The snow line moves up or down the mountain slope, depending on the orographic precipitation gradient and freezing level. Successive wedges then accumulate to give an increase in size of the snow pack with elevation (Fig. 5.7b).

Periods between storms may cause melt at lower elevations, a process that acts to steepen the overall wedge. In spring, as freezing levels rise, more precipitation falls as rain rather than snow, and increased temperatures promote more melt. Consequently, the snow wedge is eroded and finally disappears in spring or summer, with the timing of its demise dependent on elevation. For high mountains, the snow wedge never completely melts, such as in the example given by Fitzharris (1978) for the Tasman Glacier, and snow accumulates to become incorporated as névé.

There is little or no tall vegetation in the main snow areas of New Zealand, and winds are persistently strong. These factors create considerable movement and losses through sublimation or through snow being blown to lower elevations where it melts. Plumes of snow are trapped in the lee of obstructions such as tors and tall tussocks, and large drifts form downwind of terrain breaks in slope and in gullies. The result is a highly variable spatial pattern of snow deposition (Twaddle 1995).

In a study of snow accumulation on the wind-swept mountains of central Otago, Harrison (1986) found that aspect is important in controlling snow depth and water equivalent. Snow tends to accumulate in specific locations. Exposed southeast to west slopes record a water equivalent of less than 200 mm, and sheltered northeast slopes over 450 mm. Snow drifts form a useful water resource, storing approximately 4 million cubic metres in a 120 sq km catchment and contributing up to 9% of spring runoff. Snow patches continue to supply melt water for downstream irrigation well into late spring and summer. For cooler years, some can even persist into the autumn.

## PROCESSES OF SNOW MELT

The melting of snow and ice is controlled by energy transfers to the surface. Melt energy ($Q_m$) is available from several sources (Fig. 5.8), and melt can begin as soon as the snow pack or upper layers of the glacier become isothermal. In spring and summer, $Q_m$ is large, even at high elevations. Radiation energy ($Q^*$), which varies mainly with solar (short-wave) radiation, depends on the time of year, time of day and albedo (reflectivity) of the surface. New snow may have an albedo as high as 95%, but for melting snow and surfaces, albedoes are usually closer to 50%. Upward long-wave radiation normally exceeds incoming long-wave radiation, unless warm clouds or high-angle snow-free terrain are present.

Sensible heat ($Q_H$) and latent heat ($Q_E$) are transferred by turbulent motions above the surface. Their respective magnitudes depend, first, on the temperature and humidity gradients above the surface, and second, on the eddy conductivities for heat and moisture. Because melting snow or ice maintains a surface temperature of 0°C, sensible heat gain occurs when the overlying air has a positive temperature. Latent heat gain occurs when vapour pressures in the air exceed 6.11 hPa, the saturation value at 0°C, and is the release of energy as condensation occurs on the melting surface. The eddy conductivities depend mainly on wind speed, but they are also influenced in a complex way by atmospheric stability and surface roughness. As shown by Moore (1983), good estimates of the turbulent energy transfers to melting snow and ice can be obtained from measurements of wind speed, temperature and humidity at a single height above a

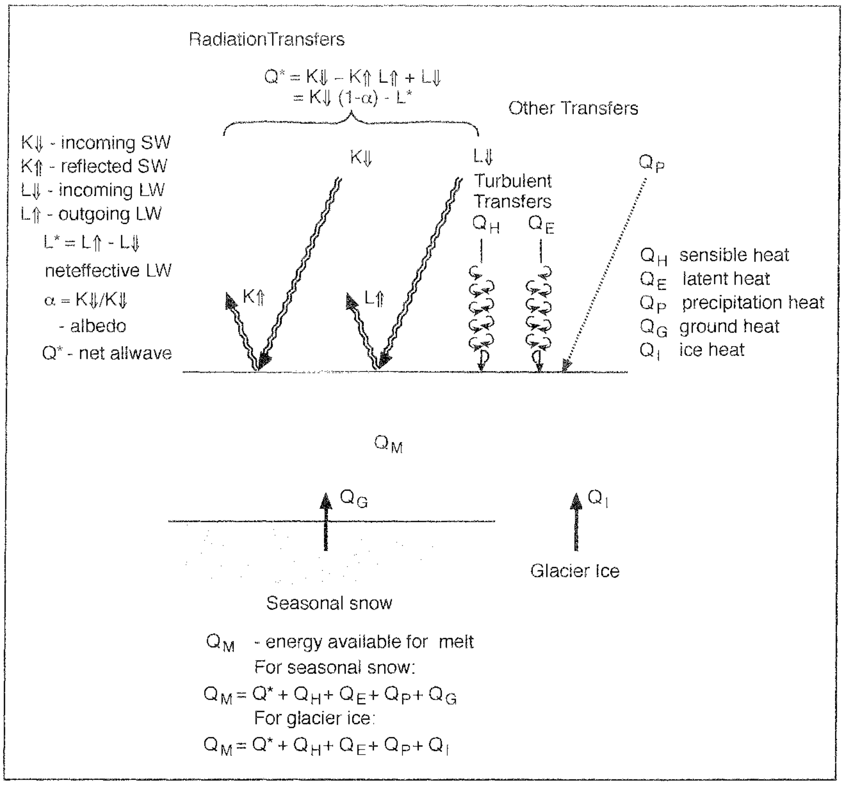


**Figure 5.8** Sources of energy for snow melt.

melting surface. The energy transfer due to rain may be estimated from rainfall rates and temperature. The energy transfer by conduction from the ground or deeper layers of a glacier may be estimated from temperature profiles.

Worldwide studies of energy transfer for melting snow show that radiation energy, $Q^*$, is usually the main source for melt, especially in forested locations and at high elevations. In New Zealand, however, the turbulent transfers of sensible and latent heat, $Q_H$ and $Q_E$, dominate because of our temperate, humid and windy conditions and nearness to the ocean. Precipitation heat transfer ($Q_P$) also makes significant, though not dominant, contributions (Table 5.2).

Lysimeter and energy balance measurements at 1750 m elevation near the Main Divide (Mueller Hut near Mount Cook village) show that radiation energy is the predominant energy source for melt of seasonal snow, but it is confined to daytime hours. During fine periods, there is a broad seasonal variation in energy supply, but nor'west storms at any time of year can produce very high turbulent and precipitation energy transfers. For example, Marcus *et al.* (1985) found that during one storm, in which 300 mm of rain fell in 10 hours on the lower Franz Josef Glacier, there was 155 mm of ice melt, even though it was mid-winter. East of the Main Divide, sensible heat is often dominant, especially during nor'westers—a frequent foehn wind in the lee of the Southern Alps (Fitzharris *et al.* 1980; Moore 1983; Neale and Fitzharris 1997). The highest elevation experiments come from the Tasman Glacier névé. Here at an elevation of 2440 m, sensible heat ($Q_H$) is the dominant source of energy for melt, with up to 78 mm/day recorded (Cutler 2002).

For many applications in hydrology, use of the energy balance approach to estimate snow melt is not very practical. Measurement of components of the energy balance requires sophisticated instrumentation in severe field environments. Estimation is possible, but uncertainty remains over key parameters, particularly exchange coefficients for sensible and latent heat ($Q_H$ and $Q_E$). Accordingly, daily snow melt is more usually estimated using the degree-day model:

$$M = f.T_{mean}$$

where

$M$ = snow melt in mm/day water equivalent
$T_{mean}$ = mean temperature for the day (°C)
$f$ = a melt factor (mm/°C/day)

This simple empirical model is found to work surprising well for catchments, provided an appropriate value of $f$ is chosen. The overseas literature considers that $f$ is between 2–8 mm/°C/day for melting snow, depending on albedo, the age of the snow and whether it has rained (Male and Gray 1981). Moore and Owens (1984) found melt factors ranging from 4–8 mm/°C/day in the Craigieburn Range. Neale (1996) reports an $f$ of 2.3 mm/°C/day at Mueller Hut at 1750 m elevation, near Mount Cook village, but some data suggests that during nor'west storms it could rise to as high as 11.5 mm/°C/day. Cutler (2002) repeated

Table 5.2 Energy sources for melting snow

| Location | Months | $Q_M$ (MJ m$^{-2}$ d$^{-1}$) | $Q^*$ (%) | $Q_H$ (%) | $Q_E$ (%) | $Q_P$ (%) | Source |
|---|---|---|---|---|---|---|---|
| Central Otago | Oct 1978 | 18.1 | 23 | 58 | 19 | <1 | Fitzharris *et al.* (1980) |
| Craigieburn (1550 m) | Nov 1976 | 8.2 | 42 | 43 | 14 | <1 | Prowse (1981) |
| | Oct 1977 | 15.1 | 33 | 64 | 3 | <1 | |
| | Oct 1979 | 13.0 | 17 | 61 | 21 | <1 | |
| | Oct 1980 | 9.2 | 27 | 60 | 12 | 1 | |
| | Average | 10.3 | 30 | 57 | 13 | <1 | |
| Temple Basin | Oct-Nov 1982 | 10.6 | 16 | 57 | 25 | 2 | Moore (1984) |
| Craigieburn Ra | Oct-Nov 1982 | 13.2 | 18 | 78 | 4 | 0 | Moore (1984) |
| Craigieburn (1750 m) | Oct-Nov 1984 | 11.8 | 60 | 16 | 14 | 10 | Pearce and Owens (1985) |
| Mueller Hut | Oct-Nov 1995 | 3.5 | 63 | 27 | 4 | 3 | Neale and Fitzharris (1997) |
| Tasman névé | Jan-Feb 2001 | 5.8 | 33 | 52 | 15 | <1 | Cutler (2002) |

his experiments, but at a higher elevation of 2440 m, on the Tasman névé and found a mean *f* of 3.4 mm/°C/day, rising to 9.1 mm/°C/day during nor'west storms.

## ROLE OF SYNOPTIC CLIMATOLOGY IN CONTROLLING SNOW AND RUNOFF

Water inflows into the storage lakes can alter much from year to year because of weather fluctuations. Research into atmospheric circulation in the vicinity of New Zealand for very high and very low inflows to hydro lakes shows that very low winter inflows are caused by a higher-than-usual incidence of westerly to southwesterly airflow (Fitzharris and Garr 1996). Such anomalies diminish runoff because in winter a higher proportion of precipitation falls as snow. For inflows to be above average in winter, higher-than-normal frequencies of northerly to northwesterly conditions are required. These bring warm rain, rather than snow, and help to melt the snow pack. Low inflows in spring and summer are produced by anomalous easterly circulations. These typically induce little rainfall near the Main Divide and are often associated with anticyclones.

The El Niño/La Niña phenomenon, as measured by the Southern Oscillation Index, appears to exert an influence on the snow hydrology. A large-scale atmospheric teleconnection exists between El Niño conditions and higher frequencies of southwesterly airstreams onto New Zealand. As a working hypothesis, early winter is colder than usual, a larger-than-normal snow pack results and hydro lake inflows are consequently reduced. By contrast, during La Niña, there are anomalous airflows over New Zealand from the northeast. Winter tends to be warmer and drier in the Southern Alps, and the snow pack is smaller. Summer inflows tend to be higher when the Southern Oscillation Index is negative in spring, and vice versa when it is positive (McKerchar *et al.* 1998).

There is a divergent circulation signal between years with high and low snow storage (de Lautour 1999). For high snow years, the winter anticyclone over Australia is larger, more intense, and it extends further south. The westerlies over New Zealand are stronger and extend further north. There is also anomalous circulation from the southwest. For low snow years, the winter anticyclone over Australia is weaker and smaller. The westerlies are weaker, and are located further south. There is anomalous anticyclonic circulation from the northeast.

## INFLUENCE OF SNOW ON FLOW OF NEW ZEALAND RIVERS

Seasonal patterns of runoff vary, depending on the relative proportions of permanent or seasonal snow. Contrasting types of distribution are illustrated in Figure 5.9. For the glaciated Hooker catchment, maximum flows occur in January, and this month alone accounts for almost 20% of the annual total. There follows a steady decrease in runoff to the winter months, when most precipitation falls as snow and there is little energy available for melt. Thus July provides only 2% of the annual flow. River flows rise again in spring and closely follow the seasonal insolation cycle, so that over half of the annual runoff occurs in the three months December to February.

By contrast, the Manuherikia catchment has no

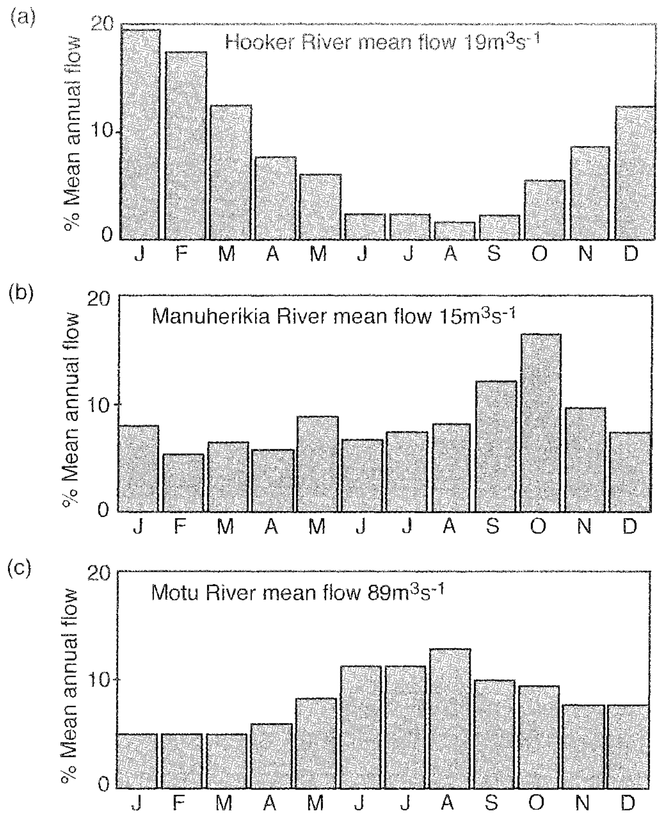


**Figure 5.9** Mean monthly runoff for contrasting catchments showing the dominant role of (a) perennial ice, (b) seasonal snow, and (c) no snow or ice.

permanent ice, but is substantially covered by seasonal snow over winter. Runoff peaks in October as temperatures rise in the spring and snow melt begins in earnest. As the area covered by snow shrinks towards summer, the flow diminishes rapidly to reach a minimum in February, and is sustained only by rare rainfalls, melt from remnant snow patches and base flow from upland bogs and groundwater.

Catchments with no snow or ice, such as the Motu, tend to produce a more even flow distribution throughout the year, with any variation being produced by seasonal changes in evaporation and rainfall. Flow tends to be higher in winter months.

More detail on the influence of seasonal snow on river flows is provided by examining the runoff patterns of tributaries of the Clutha (Jowett and Thompson 1977). Here the average altitude of the catchment is important. All rivers have low flows in July, the month with the lowest temperatures. Thence the energy available for snow melt steadily increases towards the summer solstice. The lowest catchments, such as Manuherikia, have their highest rates of snow melt and runoff in September, intermediate elevation catchments such as the Arrow and Fraser a month later, while runoff peaks in the highest catchments of Wanaka, Wakitipu, Hawea and Shotover in November.

The Fraser River shows the extreme case for a high catchment with pronounced winter snow accumulation and spring melt. Diurnal variations in river flow are clearly defined as spikes superimposed on an upward bulge in spring base flow (Fig. 5.10). The size of the spike for any given day depends on the overall synoptic weather pattern and the energy available for snow melt. On anticyclonic days, there is much solar radiation available for melt and the diurnal runoff spikes are pronounced. They grow even larger with the advent of nor'westers ahead of a cold front, when sensible flux contributes to melt, and all but disappear as cloudy, cold southerlies follow the frontal passage.

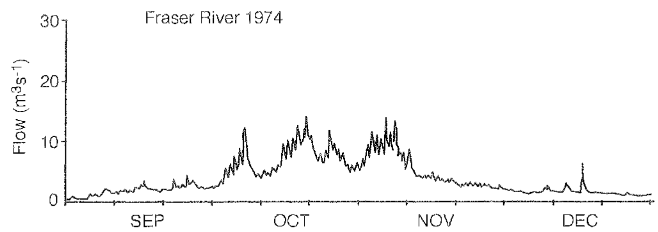


**Figure 5.10** Typical pattern of spring flow for the Fraser River, Central Otago.

## HOW IMPORTANT IS SEASONAL SNOW MELT TO RUNOFF?

Next to rainfall, the amount of water stored as snow and ice is a key factor controlling the hydrology of many mountain catchments, including those upstream of the South Island's main hydroelectricity storage lakes. Thus knowledge of these aspects is important for water management decisions. From Table 5.3 it can be seen that the average accumulated storage in these hydro catchments amounts to 15% of the annual runoff. This is predominately due to storage as seasonal snow, and represents water that is guaranteed to appear as river flow over the period October-February. It is almost the same size as controlled

**Table 5.3** Estimated long-term water balance for the main South Island hydroelectric catchments.

| Parameter | Water units (mm) | Energy units (GWh) |
|---|---|---|
| Precipitation | 2 820 | – |
| Evaporation | 550 | – |
| Runoff | 2 270 | 15 550 |
| Seasonal snow storage | 330 | 2 260 |
| Controlled lake storage | 394 | 2 700 |

Source: modified from Fitzharris (1987)

lake storage. In this context, mean seasonal snow storage is crucial in determining the seasonal timing and amount of spring and summer inflow into the hydro lakes. For the hydro catchments, spring (Oct, Nov, Dec) runoff over the period 1931-84 varied by ± 500 mm about a mean of 800 mm. Fluctuations in snow storage account for about 40% of this variation, compared with fluctuations in spring precipitation which account for 60% (Fitzharris 1987). In some smaller mountain catchments, seasonal snow melt makes larger relative contributions. For example, in the Fraser catchment of Central Otago, where runoff is used for local hydro-electricity generation, for irrigation of horticultural land, and for frost fighting, melt from snow storage provides 33% of the annual river flow on average (Fitzharris and Grimmond 1982). In many eastern mountains of the South Island, slow melt of long-lasting snow patches (Pearce and Owens 1985) can provide valuable late spring and early summer water at times of low stream flows.

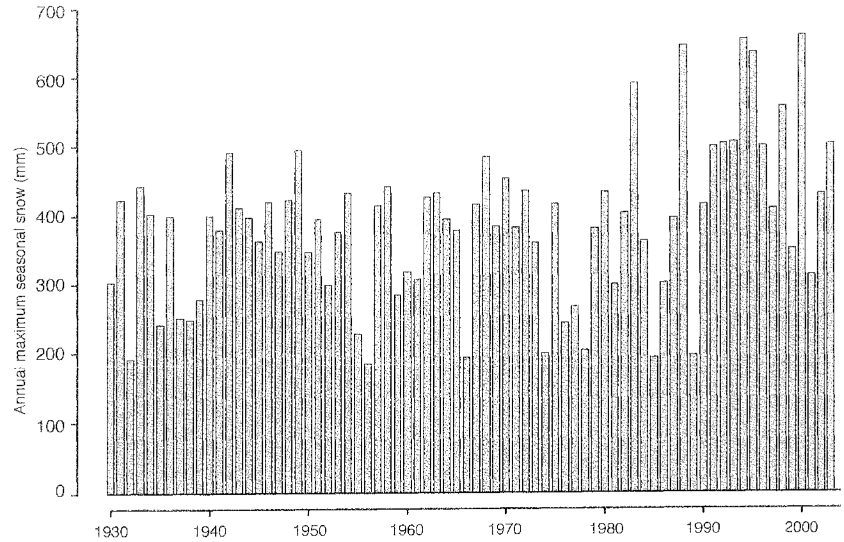


**Figure 5.11** Estimated amounts of water stored as seasonal snow in the main South Island hydro catchments 1930-2003.

Unfortunately, there are only three snow courses in New Zealand with sufficiently long records to estimate annual variability in seasonal snow storage. At Alan's Basin, at an elevation of 1750 m in the Craigieburn Range, peak snow accumulation varied between 230 mm and 1030 mm over a 12-year period, with a standard deviation of 240 mm and coefficient of variation of 44%. A 19-year record of snow pack at elevations above 1300 m in the Two Thumb Range showed a coefficient of variation of 52% (Chinn 1981). Harrison (1986) estimated snow accumulation in the Fraser basin of Central Otago over 17 winters. The driest winter gave 40% lower water equivalent and 55% less runoff than the winter with the heaviest snow cover. All these snow courses are in relatively low-precipitation areas, but demonstrate that seasonal snow storage, and hence its contribution to runoff, has a wide inter-annual range.

Fitzharris and Garr (1995) simulated daily values of seasonal snow storage for catchments of the South Island's main hydro lakes back to 1930. Seasonal snow tends to build from about May to a maximum in October. Melt is often completed by February, but water stored as seasonal snow can occasionally carry over into the following year. There is a large year-to-year variability of snow. Over 62 years, the maximum seasonal snow storage varied from less than 200 mm (1500 GWh) to over 650 mm (5000 GWh). The time series of this variability is given in Figure 5.11 and indicates considerable inter-annual variability. There is no marked trend in seasonal snow storage since 1930, although there may be larger amounts in recent years. These results confirm that the amount of water stored as snow is markedly different from one winter to the next, and as a consequence spring and summer snow-melt runoff is never consistent from one year to the next.

## SNOW MELT FLOODS

Under some conditions, rapid snow and ice melt make important contributions to floods. The role of snow melt during a major flood in the Clutha River was assessed for two subcatchments by Fitzharris *et al.* (1980). In the Fraser, where 80% of the catchment is above the normal snow line, melt provided 40% of flood runoff. Melt rates of nearly 3 mm/hr preceded rainfall. These saturated and primed the catchment, so that very quick rises in streamflow occurred when 100 mm of rain fell. In the Pomahaka, only 10% of the catchment is above the snow line, so that snow melt provided only 10 mm of the estimated 66 mm storm flow. Thus the relative contribution of snow melt to floods depends upon snow-line elevation and snow-covered area, as determined by the hypsometric curve of a catchment.

Melting of transient snow also contributes to small floods because when the snow line is lower than normal, considerable proportions of mountain catchments are involved. For instance, Moore and Prowse (1988) described a July storm that caused road washouts in the Craigieburn area. Precipitation of 123 mm in one day was augmented by over 25% by melting of a shallow snow pack at relatively low elevations. In drier mountain areas of the eastern South Island, floods can arise from snow melt alone. For example, the largest flood in one year from Camp Stream (Craigieburn Range) occurred on a day with no precipitation, but with foehn nor'west conditions producing strong melt (Moore and Prowse 1988).

## CAN SNOW SURVIVE FUTURE CLIMATE CHANGE?

Future changes in water resources, electricity supply and electricity demand are likely if global warming occurs. This is because snow and glaciers are very sensitive to changes in temperature. The sensitivity of snow storage to changes in precipitation and temperature is shown as a response surface for South Island hydro catchments in Figure 5.12. Snow storage declines as temperature increases and precipitation decreases, but the curved isolines indicate that the relationship is not linear. For example, with temperatures colder than present, snow storage becomes more sensitive to changes in precipitation. Climate warmings envisaged by two plausible scenarios produce much reduced snow storage: 170 mm for warmer and wetter Scenario 1 (compared with 330 mm at present); and about 50 mm for warmer and drier Scenario 2 (Fig. 5.12). Seasonal snow storage is so diminished in Scenario 2 that the seasonal distribution of runoff will be greatly affected.

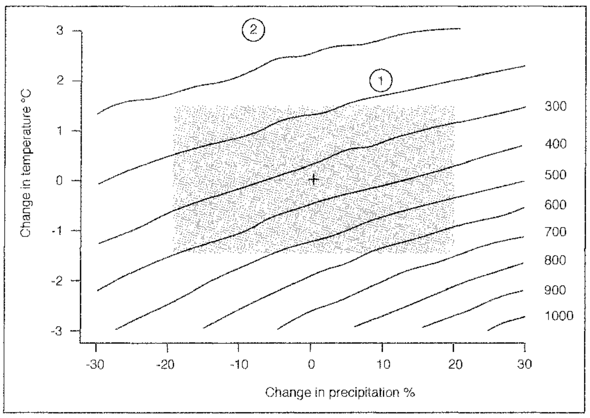


**Figure 5.12** Impact of climate change on water stored as seasonal snow in the main South Island hydro catchments. Diagram is a response surface, which shows the sensitivity of snow with respect to changes in temperature and precipitation. Snow is given as mm water equivalent. Shaded area shows range of inter-annual variablity for present climate.

Assuming increases in temperature of 3°C and precipitation of 15%, Fitzharris (1989) suggested a rise in snow line of 300–400 m, a decrease in snow accumulation below 2300 m and a reduction in winter snow-covered area for South Island hydro catchments from 45% to 28%. These changes will markedly alter the flow regime of many South Island rivers. Modelling studies by Fitzharris and Garr (1996) suggest an increase in inflow to hydro storage lakes of 40% in winter, but a decrease of 13% in summer. Annual runoff would increase by 14%. All these changes favour increased hydro generation and reduce the demand for water storage. Such forecasts are preliminary because current year-to-year variability in snow storage is larger than the simulated changes in the mean, and because regional estimates of future climate are at present rather crude.

## REFERENCES


Barringer, J.R.F. 1989: A variable lapse rate snowline model for the Remarkables, Central Otago, New Zealand. Journal of Hydrology (NZ) 28(1): 32-46.

Chinn, T.J. 1969: Snow survey techniques in the Waitaki Catchment, South Canterbury. Journal of Hydrology (NZ) 8(2): 68-76.

Chinn, T.J. 1981: Snowfall variations, hazards and snow melt. In: Mountain Lands Workshop, Water and Soil Science Centre, Christchurch. Ministry of Works and Development Report No. WS 525, 1-21.

Christchurch Engineering Lifelines Group 1997: Risks & Realities: A Multi-disciplinary Approach to the Vulnerability of Lifelines to Natural Hazards. Centre for Advanced Engineering, University of Canterbury, Christchurch.

Colbeck, S.; Akitaya, E.; Armstrong, R.; Gubler, H.; Lafeuille, J.; Lied, K.; McClung, D.; Morris, E. 1990: The International Classification for Seasonal Snow on the Ground, International Commission for Snow and Ice of the I.A.S.H.

Cutler, E.S. 2002: High elevation seasonal snow melt at the Tasman Glacier névé, Southern Alps, New Zealand. MSc thesis, University of Otago, Dunedin.

de Lautour, S.M.A. 1999: Seasonal snow storage in hydro catchments and atmospheric circulation patterns. MSc thesis, University of Otago, Dunedin.

Fitzharris, B.B. 1978: Problems in estimating snow accumulation with elevation on New Zealand Mountains. Journal of Hydrology (NZ) 17(2): 78-90.

Fitzharris, B.B. 1979: Snow Hydrology. In: Murray, D.L.; Ackroyd, P. (eds.), Physical Hydrology – New Zealand Experience, New Zealand Hydrological Society, Wellington, 2-43.

Fitzharris, B.B. 1987: A method for indexing the variability of alpine snow cover over large areas. In: Large Scale Effects of Seasonal Snow cover, Proc. Vancouver Symposium, I.A.S.H. Pub. No.166, 139-150.

Fitzharris, B.B. 1989: Climatic change and the future of New Zealand's snow cover and snowline. In: Climate Change, the New Zealand Response. Proceedings of a workshop held in Wellington, New Zealand. Ministry for the Environment, 253-254.

Fitzharris, B.B.; Garr, C.E. 1995: Simulation of past variability in seasonal snow in the Southern Alps, New Zealand. Annals of Glaciology 21, 377-382.

Fitzharris, B.B.; Garr, C.E. 1996: Climate, Water Resources and Electricity. In: Bouma, W.J.; Pearman, G.I.; Manning, M.R. (eds.), Greenhouse: Coping with Climate Change, CSIRO Publishing, Collingwood, 263-280.

FITZHARRIS, B.B.; GRIMMOND, C.S.B. 1982: Assessing snow storage and melt in a New Zealand mountain environment. In: Hydrological Aspects of Alpine and High Mountain Areas, Proc. Exeter Symposium, July 1982, I.A.H.S. Pub. No.138, 161-168.

FITZHARRIS, B.B.; LAWSON, W.; OWENS, I. 1999: Research on glaciers and snow in New Zealand. Progress in Physical Geography 23 (4): 469-500.

FITZHARRIS, B.B.; MCALEVEY, B.P. 1999: Remote sensing of seasonal snow cover in the mountains of New Zealand using satellite imagery. Geocarto International 14(3): 33-42.

FITZHARRIS, B.B.; MCGANN, R. 1989: Mapping Seasonal Snow Cover from Weather Satellites. In: Welch, R. (ed.), Proc. Fifteenth New Zealand Geography Conference, 235-238.

FITZHARRIS, B.B.; OWENS, I.; CHINN, T. 1992: Snow and glacier hydrology. In: Mosley, M.P. (ed.), Waters of New Zealand. New Zealand Hydrological Society, Christchurch, 5: 75-93.

FITZHARRIS, B.B.; STEWART, D.; HARRISON, W. 1980: Contribution of snow melt to the October 1978 flood of the Pomohaka and Frazer Rivers, Otago. Journal of Hydrology (NZ) 19: 84-93.

HARRISON, W. 1986: Seasonal accumulation and loss of snow from a block mountain catchment in Central Otago. Journal of Hydrology (NZ) 25(1): 1-17.

HICKMAN, J.S. 1972: Estimates of snowline from meteorological satellite pictures 1968-71. Unpub. New Zealand Meteorological Service Res. Report.

HUGHES, J.C. 1974: The snow of August 1973: Tussock Grasslands and Mountain Lands Institute Review 30: 64-100.

JOWETT, I.G.; THOMPSON, R.D. 1977: Clutha Power Development: flows and design floods (Interim Report). Power Division, Ministry of Works.

MALE, D.M.; GRAY, D.H. 1981: Snowcover, ablation and runoff. In: Gray, D.H.; Male, D.M.(eds.), Handbook of Snow – Principles, processes, management and use, Pergamon Press, Toronto, 360-436.

MARCUS, M.G.; MOORE, R.D.; OWENS, I.F. 1985: Short-term estimates of surface energy and ablation over the lower Franz Josef Glacier, South Westland, New Zealand. New Zealand Journal of Geology and Geophysics 28: 559-567.

MCALEVEY, B.P. 1998: A distributed seasonal snow model for New Zealand. MSc thesis, University of Otago, Dunedin.

MCGREGOR, G.R. 1990: Snowpack structure and avalanching, Craigieburn Range, New Zealand. New Zealand Journal of Geology and Geophysics 33: 405-417.

MCKERCHAR, A.I.; PEARSON, C.P.; FITZHARRIS, B.B. 1998: Dependency of summer lake inflows and precipitation on spring SOI. Journal of Hydrology 205: 66-80.

MCNULTY, D.N.; FITZHARRIS, B.B. 1980: Winter avalanche activity and weather in a Canterbury alpine basin. New Zealand Journal of Geology and Geophysics 23: 103-111.

MOORE, R.D. 1983: A Comparison of the Snowmelt Energy Budgets in Two Alpine Basins. Arch. for Met. Geophys. and Bioclimat. B33:1-10.

MOORE, R.D. 1984: Snow hydrology of a mountainous rain-on-snow environment – the Waimakariri catchment, New Zealand. PhD thesis, University of Canterbury, Christchurch.

MOORE, R.D.; OWENS, I.F. 1984: Modelling alpine snow accumulation and ablation using daily climate observations. Journal of Hydrology (NZ) 23: 73-83.

MOORE, R.D.; PROWSE, T.D. 1988: Snow hydrology of the Waimakariri catchment, South Island, New Zealand. Journal of Hydrology (NZ) 27: 44-68.

NEALE, S.M. 1996: Measuring and modeling snow melt in a New Zealand Alpine environment. MSc thesis, University of Otago, Dunedin.

NEALE, S.M.; FITZHARRIS, B.B. 1997: Energy balance and synoptic climatology of a melting snow pack in the Southern Alps, New Zealand. International Journal of Climatology 17: 1595-1609.

O'LOUGHLIN, C.L. 1969: Further snow investigations in the Craigieburn Range. New Zealand Forest Service, Protection Forestry Report 52.

PATERSON, W.S.B. 1969: The physics of glaciers. Pergamon Press, Oxford.

PEARCE, H.M.; OWENS, I.F. 1985: Ablation of a late-lying snowpatch, Alans Basin, Craigieburn Range, New Zealand. Proceedings New Zealand Hydrological Symposium, 1985, 45-46.

PROWSE, T.D. 1981: The Snow Environment of the Craigieburn Range. PhD thesis, University of Canterbury, Christchurch.

PROWSE, T.D.; OWENS, I.F. 1984: Characteristics of snowfalls, snow metamorphism and snowpack structure with implications for avalanching, Craigieburn Range, New Zealand. Arctic and Alpine Research 16: 107-118.

SELIGMAN, G. 1936: Snow structure and ski fields. Macmillan, London.

TECHNICAL SUBCOMMITTEE ON SNOW, N.Z. Committee for the I.H.D. 1969: Preparatory report. Water and Soil Division Ministry of Works.

THOMAS, I.L.; LEWIS, A.J.; CHING, N.P. 1978: Snowfield assessment from Landsat. Photogrammetric Engineering and Remote Sensing 44: 493-502.

TWADDLE D.G. 1995: Modeling snow cover in short tussock grassland. MSc thesis, University of Otago, Dunedin.

WEIR, P.L.; OWENS I.F. 1981: Snowpack structure at Mt Hutt, Canterbury in 1979. New Zealand Journal of Science 24: 95-102.