Journal of Hydrology (N.Z.) Vol. 27, No. 1, 1988

SNOW HYDROLOGY OF THE WAIMAKARIRI CATCHMENT, SOUTH ISLAND, NEW ZEALAND

R. D. Moore

Geography Department, McGill University Montréal, P.Q., Canada

T. D. Prowse

National Hydrology Research Centre, Environment Canada, Saskatoon, Canada

ABSTRACT

The Waimakariri River drains eastward from the main divide of New Zealand's South Island. Most streamflow derives from the mountainous upper catchment, which is typical of other major catchments in the Canterbury region. Snow accumulation and melt processes are governed by interactions between rapidly changing weather conditions and the steep terrain. Fluctuations in freezing level during storms produce complex spatial and temporal variations of accumulation and melt, resulting in substantial sampling problems. Research and data collection have been hampered by poor access, lack of good stream gauging sites and a paucity of climate data from the mountainous regions. The common coincidence of heavy rainfall with snowmelt complicates the application of water balance methods.

Estimates of mean seasonal snowmelt volumes range from nine to fifteen percent of mean annual streamflow. However, snow accumulation at the end of winter can be least twice the mean in some years. Snowmelt contributions to floods cannot be accurately assessed with currently-available information, but evidence is presented to indicate that snowmelt and rain-on-snow events play a role in the flood hydrology of basins in the eastern portions of the catchment, where rainfall intensities are lower than in the areas closer to the main divide.

INTRODUCTION

Most of the streamflow in the South Island of New Zealand derives from the mountains, where varying portions of precipitation fall as snow. In the block mountains of Central Otago, the magnitude and timing of snow accumulation and melt and the contributions of snowmelt to streamflow have been documented by Fitzharris and Grimmond (1982), Fitzharris et al. (1980) and Harrison (1986a,b). Knowledge of snow hydrology in the Fraser River catchment in Central Otago, where snow survey and snowpillow data have been collected, suffices for many water management decisions. The hydrological role of snow in the headwater catchments of Canterbury rivers has been less well documented, although water balance estimates indicate that snowmelt may be hydrologically significant. For example, Anderton (1974) estimated that twenty percent of the inflows to Lake Pukaki in the Waitaki catchment result from the melting of seasonal snow cover, and Bowden

(1983) estimated that snowmelt contributes on average about twelve percent of the mean annual streamflow in the Rakaia River.

Some attention has been paid to snow hydrology in the Waitaki and Rakaia catchments (e.g. Anderton, 1974; Bowden, 1983; Chinn 1969, 1981), but most research and data collection related to snow hydrology in the Canterbury mountains have been carried out in the Waimakariri catchment (see Fig. 1). This catchment contains one of only three roads which cross the Southern Alps, thus affording access to the mountainous headwaters. In addition, a number of huts operated by ski clubs provide convenient accommodation for conducting research above the snowline.

This paper summarises what is known about the snow hydrology of the Waimakariri catchment. The first three sections outline the geographic setting, streamflow regime and available sources of data about the Waimakariri catchment, followed by a review of published and unpublished work on processes governing snow accumulation, melt and runoff. Attempts to monitor and model snow storage and melt are then reviewed, and an attempt is made to quantify the hydrological significance of snow in the Waimakariri catchment, followed by suggestions for future work.

GEOGRAPHICAL SETTING OF THE WAIMAKARIRI CATCHMENT

The Waimakariri River heads on the main divide of the Southern Alps in the central South Island and drains from mountain basins and through a gorge before crossing the Canterbury Plains to the Pacific Ocean (Fig. 1). After the river leaves the foothills gorge, between five and twenty-five percent of its flow is lost to ground water as it crosses the Canterbury Plains (Dalmer, 1971). The upper catchments of Canterbury rivers (from the Hurunui to the Waitaki, see Fig. 1) are similar in terms of geology, vegetation and climate, although catchments to the south are higher and more heavily glacierized.

Ridgetop and peak elevations range from 1500 to 2400 m, while valley bottoms lie at about 600 to 800 m. The highest summits and steepest slopes are located close to the main divide, with the eastern mountains and foothills having rounded ridge tops and valley side slopes between 30° and 40° above 1200 m. The hypsometry of the 2340 km^2 catchment above the gorge is shown in Fig. 2. The bedrock geology is dominated by strongly indurated and folded sandstones and siltstones of the Torlesse Supergroup. This rock erodes readily and incised streams and gorges are common, particularly in the higher rainfall areas close to the main divide. Extensive tracts of the mountain slopes are mantled by scree deposits. Streams tend to be boulder torrents in their highest reaches, but broaden out to an unstable "braided" habit downstream.

The most recent glaciation ended ca. 15000 B.P. Pleistocene glaciers extended from all main divide valleys through the gorge to the edge of the plains. Fluvial and colluvial processes have extensively modified much of the post-glacial landscape. Present glaciation is restricted to forty-nine remnant cirque glaciers covering an area of 4.3 km^2 along the eastern flanks of the main divide (Anderton, 1973). The climatic snowline elevation is presently 2150 m (Burrows, 1977) and the glacial snowline is at 1900 m (Chinn and Whitehouse, 1980).

Forests composed of several species of beech (*Nothofagus spp.*) extend

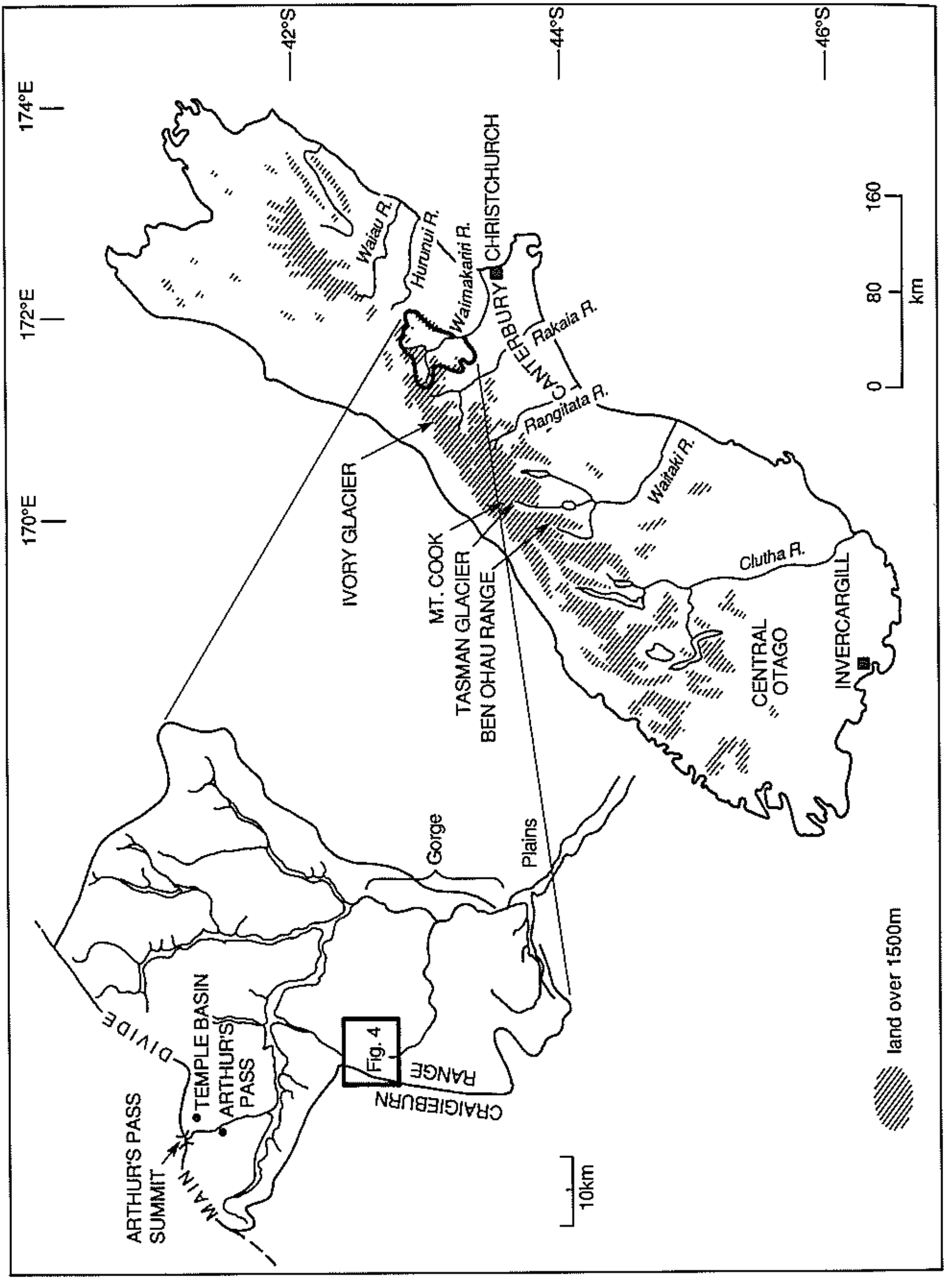


FIG. 1—Map of the South Island, New Zealand, and the upper Waimakariri catchment. Location of Craigieburn Range sites shown by square.

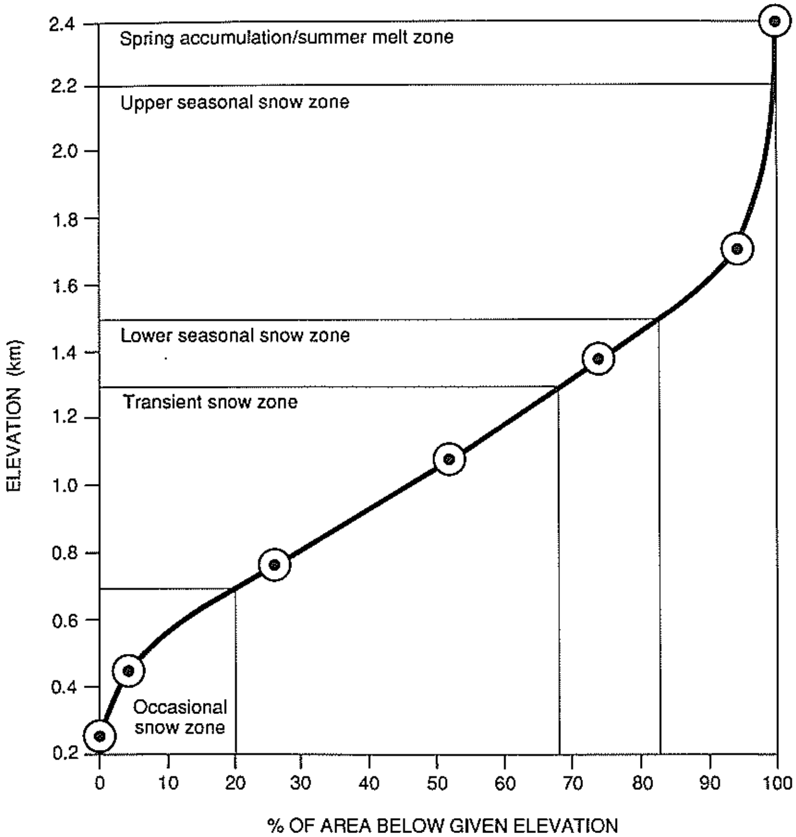


FIG. 2—Hypsometry of the upper Waimakariri catchment (after Hayward, 1967). Circles indicate contour measurement points.

up to 1370 m. Fires occurring before European intrusion cleared extensive areas of forest; over half of the area below treeline is currently covered by grassland (Hayward, 1967). Alpine grassland and bare scree dominate areas above treeline.

The South Island lies in the latitudes known as the "roaring forties." Cyclonic, anticyclonic and frontal systems, usually moving from west to east over the Tasman Sea, dominate weather, producing a typical "west coast mid-latitude" climate (Fitzharris, 1978). Mean annual precipitation decreases from over 8000 mm along the main divide to less than 1000 mm in the eastern portions of the catchment, over a distance of about 30 km (North Canterbury Catchment Board and Regional Water Board [NCCBRWB], 1986). Monthly precipitation normals in the mountains show two peaks, the main one in spring with a lesser one in autumn, which coincide with seasonal strengthening of the westerly circulation. These peaks are not as well defined at stations

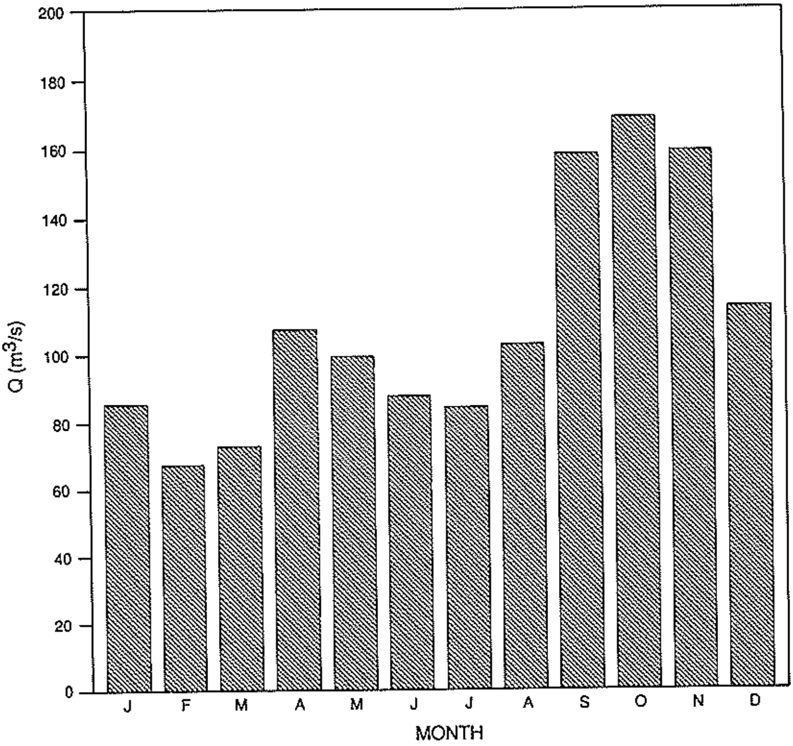


FIG. 3—Mean monthly streamflow, Waimakariri River.

in the eastern portion of the catchment because these areas are influenced by southerly and easterly weather. The Craigieburn Range, where many of the relevant studies have been carried out, lies three ranges in the lee of the main divide, and is less influenced by westerly weather systems than areas like the Temple Basin site, nearer the main divide (Fig. 1).

STREAMFLOW REGIME OF THE WAIMAKARIRI RIVER

The most reliable gauging site on the Waimakariri River is near its mouth, so measured flows are affected by losses to the Canterbury Plains. Mean annual runoff from the catchment above the gorge is approximately 1600 mm with a coefficient of variation of 0.21; monthly flows are more variable, with coefficients of variation between 0.33 and 0.55 (NCCBRWB, 1986). The flow regime (see Fig. 3) is marked by a peak in the spring months which coincides with the seasonal peak in precipitation along the main divide, but is believed to be augmented by snowmelt (Bowden, 1977). A secondary peak coincides with the autumn precipitation maximum in the western part of the catchment. Low flows occur in February, when precipitation is low and evapotranspiration loss is high, and in July, when a portion of precipitation is stored as snow.

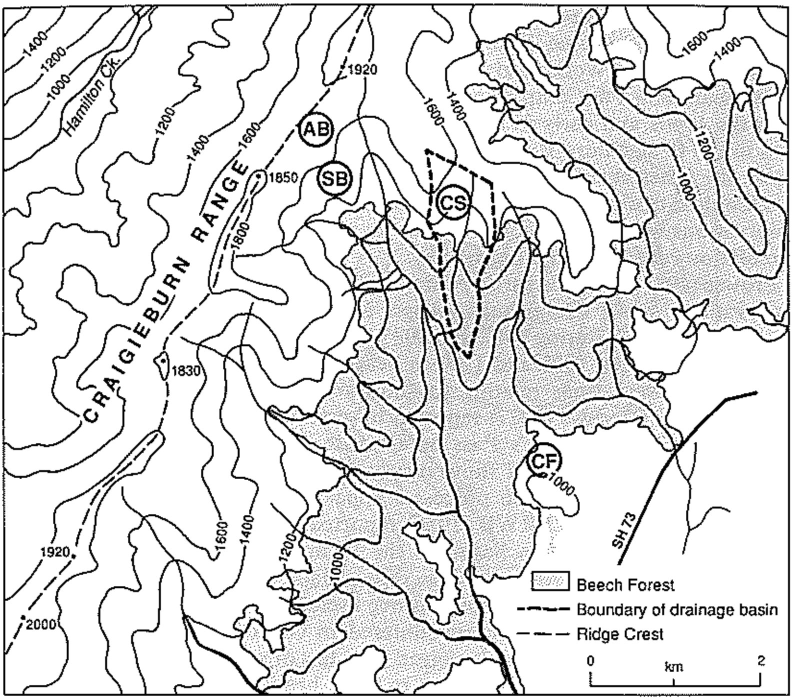


FIG. 4—Location of Craigieburn Range sites.

CF — Craigieburn Forest	SB — Ski Basin
CS — Camp Stream	SH73 — State Highway 73
AB — Alan's Basin	

Floods can occur at any time of the year, but approximately sixty percent of all floods occur from October through February (Hayward, 1967). Most floods are associated with northwesterly storms (NCCBRWB, 1986).

DATA BASE

The New Zealand Meteorological Service (NZMS) and NCCBRWB maintain a network of temperature and precipitation stations in and surrounding the Waimakariri catchment; the highest of these is at Arthur's Pass, at an elevation of 740 m. The New Zealand Forest Service (NZFS) established a number of climate stations on the eastern side of the Craigieburn Range, of which two are currently maintained: Ski Basin at an elevation of 1550 m, and Craigieburn Forest at 914 m. Ski Basin is a southeast facing cirque ranging between 1500 and 1800 m elevation. Measurements at these stations include solar radiation, temperature, humidity and wind run. In addition, weather and snowpack observations are made during the ski season at several sites in the Craigieburn Range as part of routine avalanche control operations. Locations of the Craigieburn Range sites are shown in Fig. 4.

A stream gauging installation on Camp Stream, which drains a first order catchment of 0.94 km^2 in the Craigieburn Range, was set up by the NZFS during the International Hydrological Decade and is currently maintained by the Ministry of Works and Development. A recording precipitation gauge is also maintained at Camp Stream.

In the 1960s and early 1970s, the NZFS carried out snow surveys at two sites in the Craigieburn Range: one at 1450 m elevation within Camp Stream basin, the other at 1750 m in Alan's Basin, a cirque basin between Ski Basin and Camp Stream (Morris and O'Loughlin, 1965). A range of other snow observations were made, including density profiles, liquid water contents and percolation rates (O'Loughlin, 1969).

Prowse (1981) collected further climate data and made systematic snowpack observations in the Craigieburn Range. Moore (1984) maintained a micrometeorological site during 1982 and 1983 at 1450 m in Temple Basin, a cirque basin lying between 1300 and 1900 m near the main divide (see Fig. 1), and also made observations relating to snowmelt runoff processes during two snow seasons. Burrows (1977) reported observations of snow properties and distribution in the Arthur's Pass area, made as part of ecological studies. Chinn (1969, 1981) reported snow survey data from the Waitaki catchment, in which snow accumulation processes appear to be similar to those in the Waimakariri. The NCCBRWB (1986) reported data on streamflows from a number of small mountain basins in the Waimakariri catchment, including the Bealey River near Arthur's Pass. These data have not been analysed in terms of snow hydrology, but could prove useful.

A number of problems are encountered in collecting data in the South Island mountains. The terrain is rugged with few or no roads, requiring either helicopter support or access on foot to most headwater locations. Very few stable, accessible gauging sites are to be found on streams, so streamflow data are often collected only for short periods for specific purposes. Winds are frequent and strong, necessitating robust instrumentation; in addition, riming during winter is common. Finally, mountain parrots called "keas" (*Nestor notabilis*) can wreak havoc with instrumentation, even in the presence of an operator (see for example Anderton and Chinn, 1978; Harding, 1972; Marcus et al., 1985).

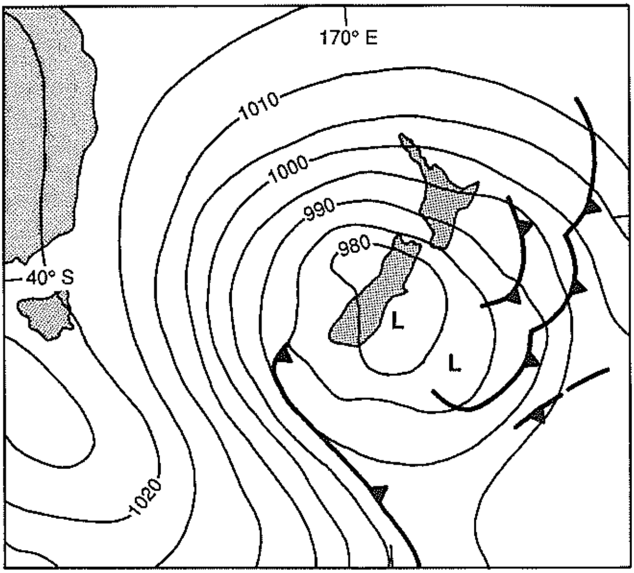
SNOW DEPOSITION

Lowland Snow Storms

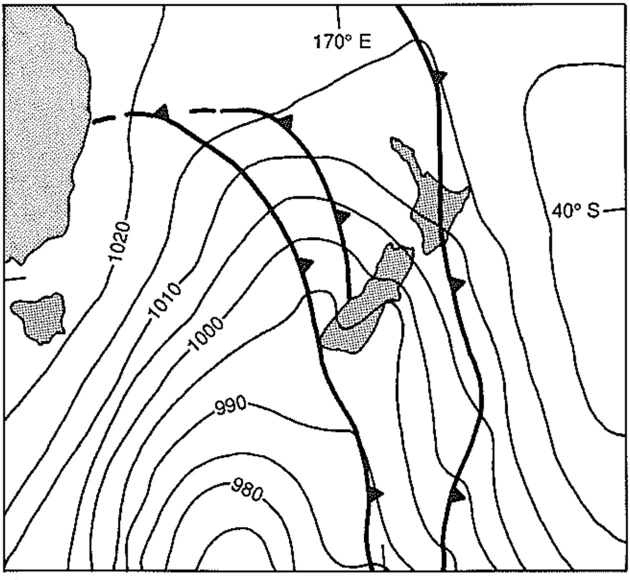
Snow storms affecting the lowlands of the South Island have been studied by Neale and Thompson (1977) using ground and radiosonde data. Following Younkin (1968), storms were considered to be of the warm advection type when the atmospheric structure was similar to that of warm fronts found in storms affecting the western United States, and were classified as cold vorticity storms when large scale vorticity advection prevailed, usually in broad flows of unstable cold air from high latitudes. Such conditions are typical of "polar low" storms which are important snow producers over Britain, for example (Lyall, 1972).

Alpine Snow Storms

Although some of the same features of the lowland storms might apply to the alpine zone, a detailed examination of the major storms which deposit



September 13, 1976 - 0000 GMT



June 9, 1976 - 1200 GMT

FIG. 5—Synoptic charts of the two types of snow storms affecting the Craigieburn Range. Lower chart is 5a: upper chart is 5b.

snow in the mountains of the South Island is difficult because of the scarcity of high-elevation climate data. A number of researchers have, however, pointed to the importance of the passage of meridional troughs of low pressure in producing large snowfalls over the Craigieburn Range (O'Loughlin, 1969; Prowse, 1981; McGregor, 1984). Fig. 5a depicts an example of this storm type which is usually characterised by falling temperatures. Warm air at the beginning of the storm commonly belongs to a north to northwesterly airflow preceding the trough and frontal edge; it is subsequently replaced by cooler southerly winds behind the front.

Within the trough of many of the severest storms, there is a centre of low pressure which passes over or in close proximity to the South Island (Prowse 1981). Although the eastward progression of these centres is usually rapid, some become stalled along the east coast of New Zealand by large blocking anticyclones (Fig. 5b). Snowfall is then received primarily from south to northeast airflows. According to O'Loughlin (1969), this is the second major type of storm which affects the Craigieburn Range.

As suggested by Griffiths and McSaveney (1983), in a east-west transect study, westerly storms (W-NW) bring maximum precipitation to the western portions of the Southern Alps peaking near the crest of the nearest-seaward ranges or main divide. Then, partly as the result of a foehn effect, westerly-derived precipitation decreases with increasing distance from the western coast. In contrast, southerly and easterly wind flows (NE-S) bring maximum precipitation to the eastern portion of the catchment. Westward penetration of such storms is also believed to be limited, explaining Burrows' (1977) observation that almost all snow received near the main divide derives from westerly storms.

Freezing levels

Using meteorological data recorded at the Craigieburn Forest and Ski Basin stations, Prowse (1981) estimated the elevations of storm snowlines from a five-year record of eighty-four snowstorms (defined as storms producing at least 5 mm water equivalent of snow at Ski Basin). During the main winter months, snow storm freezing levels frequently fall to 700 to 900 m, although most precipitation is deposited while freezing levels are in the 1100 to 1300 m range. Combinations of precipitation types are common in this middle zone, with first rain, and then sleet and snow, falling as the freezing level descends, primarily because of the shift in wind direction. Christchurch radiosonde data show that freezing levels during southwesterly airflows are typically several hundred metres lower than during northwesterly airflows (Prowse, 1981). Above approximately 1500 m snow usually dominates, although higher elevations can also experience above-freezing temperatures during storms: approximately half of the eighty-four snowstorms were characterised by maximum temperatures above 0°C at Ski Basin.

Wind effects

Snow is subject to severe redistribution due to strong winds, especially near ridge tops. Rowe (1968) reported wind gusts in excess of 67 m/s along the summit of the Craigieburn Range, and McSaveney (1978) estimated that a daily average wind speed of 50 m/s has a return period of one year on

Mount Hutt, which lies just outside the Waimakariri catchment. Avalanches can also significantly affect snow distribution, but on a more local scale (Burrows, 1977; Weir, 1979).

Forest effects

O'Loughlin (1969) found that the accumulated snowpack under beech forest was usually thirty to forty percent less than that in adjacent open areas and suggested that evaporation of intercepted snow could account for the difference. Another possibility is that much of the snow intercepted in the tree crowns melted from exposure to the above-freezing temperatures which often prevail during snowstorms below treeline, and subsequently dripped through the canopy and snow pack to the underlying soil. This process is consistent with snow densities under forest being greater than those in the open, which would result from increased contact with liquid water (Morris and O'Loughlin, 1965).

SNOWMELT CHARACTERISTICS

Snowmelt rates depend upon both the rate of energy exchange across the snowpack interfaces and the thermal state of the pack. In the Waimakariri catchment, measurements have shown that snowpacks below about 2000 m are usually at or near the melting point throughout winter, and heat deficits rarely exceed the energy required to melt 6 mm water equivalent (Prowse, 1981; Moore, 1984). Conductive heat flux at the base of the snowpack was found to be negligible as a source of energy for melt at Ski Basin (Prowse and Owens, 1982), although it is important for preventing ground freezing beneath the snow cover (Burrows, 1977). Rates of energy exchange at the upper surface are thus the primary controls on melt rates in the Waimakariri catchment.

Net radiation is relatively unimportant as an energy source for melt during winter, except on north-facing slopes, and even then it can be offset by sensible heat loss when southerly winds are blowing (Prowse, 1981; Moore, 1984). Turbulent exchange of sensible and latent heat is efficient because most of the Waimakariri catchment is unforested and snow cover is exposed to the prevailing strong winds. During storms, snow cover at sites below the freezing level will gain sensible and possibly latent heat from the atmosphere, favouring melt, while snow cover at sites above the freezing level will tend not to melt because of heat loss to the atmosphere. Snowfall at higher elevations often coincides with rain and melt at lower elevations. The freezing level during storms fluctuates in the zone below 1500 m during winter, so this zone is subject to periodic melt, leading to thin snow cover. Only occasional winter storms involve freezing levels higher than 1500 m, resulting in less frequent winter melt.

Energy for turbulent exchange can be derived either through regional advection associated with air mass movements and foehn wind effects or through local advection from snow-free areas. In the Waimakariri catchment, lower slopes in spring are normally free of snow. Radiative heating of these bare slopes on clear, calm days warms the air and can generate upslope winds which enhance turbulent exchange over adjacent snowfields. Moore and Owens (1984a) recorded this phenomenon at Temple Basin, but found that the most

TABLE 1—Mean melt rates (M) and energy exchanges estimated for the Ski Basin (SB) and Temple Basin (TB) sites. M is in mm/d and all energy fluxes are in MJ/m²/d).

M	Q*	QH	QE	QP	Notes
3	-4.5	4.7	0.6	0.1	(1)
33	3.3	6.4	1.4	0.1	(2)
40	2.4	10.3	0.5	0.0	(3)
32	1.7	6.0	2.7	0.2	(4)

Key: Q* — net radiation
QH — atmospheric sensible heat
QE — atmospheric latent heat
QP — sensible heat of rain

Notes:
(1) average of 20 days of winter melt events at SB between 1976 and 1980.
(2) average of 33 days of spring melt events at SB between 1976 and 1980.
(3) average of 12 days of spring melt at SB 28 October 1982 — 8 November 1982.
(4) average of 12 days of spring melt at TB 28 October 1982 — 8 November 1982.

rapid melt coincided with storms during which insolation and local advection were minimal and melt was dominated by regional advection. Estimated snow surface energy exhanges and melt rates at Ski Basin and Temple Basin have been presented by Prowse (1981), Prowse and Owens (1982), Moore (1983) and Moore and Owens (1984a), and are summarised in Table 1. These results show the dominance of atmospheric sensible heat as a source of energy, reflecting the exposed nature of the sites and the influence of regional advection. The differences between the two sites arise from the greater maritime influence at Temple Basin during westerly weather: greater cloud cover, cooler more humid air and more rain.

Some of the winter melt events investigated by Prowse (1981) contained non-melt periods, so the mean melt rate was low. However, the estimated net energy exchange on the day of maximum winter melt would have produced 64 mm water equivalent of melt, with the greatest contribution from atmospheric sensible heat. Most of the winter melt events coincided with rain. The greatest estimated daily spring melt totals were 64 mm/d at Temple Basin and 93 mm/d at Ski Basin. The maximum daily melt at Ski Basin occurred during foehn conditions. No similar work appears to have been carried out on snowmelt under beech forest or at any sites below the seasonal snowline.

The onset of the main spring thaw usually occurs between mid-September and mid-October (see Fig. 6), during the seasonal strengthening of the westerly circulation. The timing of onset is related to whether the westerly systems in early spring bring rain or snow, which is related to their cyclonicity. Cyclonic systems (as in Fig. 7a) drawn air from the south and bring snow, while anticyclonic circulation (as in Fig. 7b) draws warm, moist air from the northwest, bringing rain and producing melt.

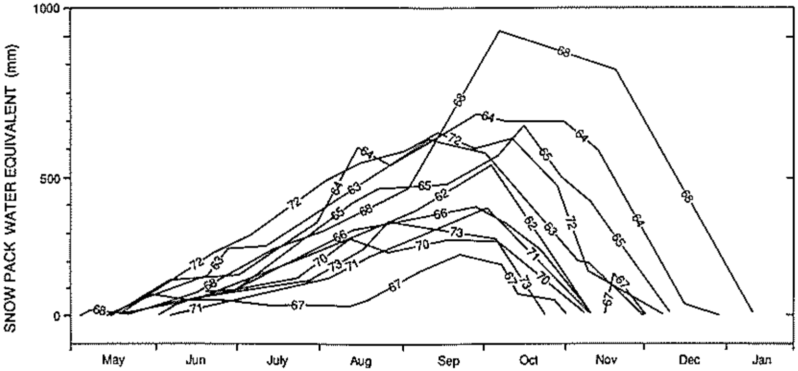


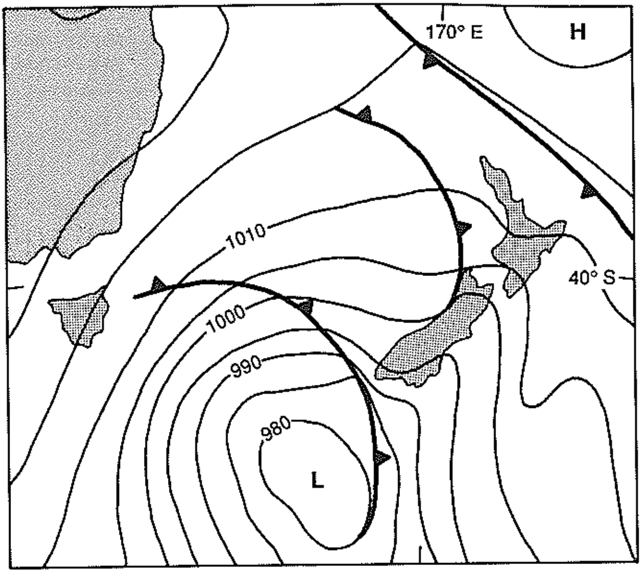
FIG. 6—Snow course measurements at Alan's Basin (updated from O'Loughlin 1969).

VARIATIONS IN SNOW ACCUMULATION

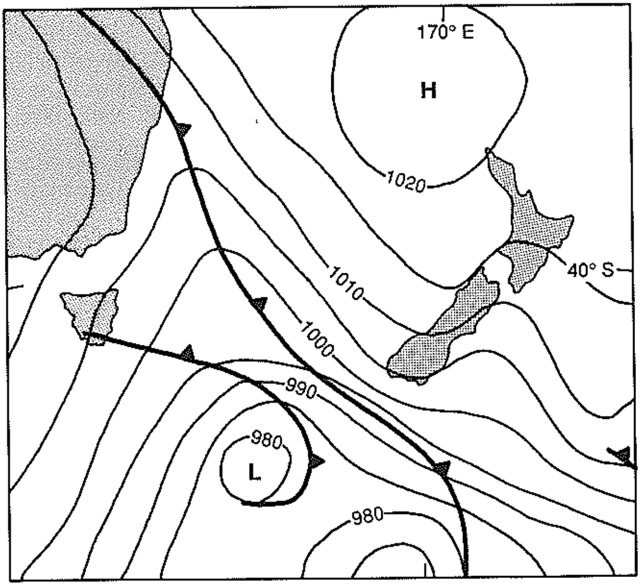
Based on altitudinal variations in deposition and melt, and on observations by O'Loughlin (1969), Burrows (1977), Chinn (1969; 1981), Owens et al. (1983) and the present authors, the Waimakariri catchment can be separated into five elevation zones having distinctive accumulation patterns. This division represents a refinement of the snow region classification of the New Zealand National Committee for the I.H.D. (1969), which distinguished snow free, occasional snow cover, seasonal snow cover and permanent snow and ice zones. Descriptions of the five zones follow. (See also Fig. 2.)

1) Below 700 m large snowfalls usually occur less than once per year (occasional snow zone).
2) Between 700 and 1300 m, snow cover is transient, and the snowline fluctuates throughout winter with the alternation of snowfall and snowmelt (transient snow zone).
3) Between 1300 and 1500 m, snow cover is usually continuous both spatially and temporally — except possibly at windswept and/or north facing sites — but mid-winter melt significantly affects snow accumulation in this zone, maintaining relatively thin, warm snowpacks (lower seasonal snow zone).
4) Mid-winter melt events are less important above 1500 m, where only occasional events produce marked ablation (Morris and O'Loughlin, 1965) (upper seasonal snow zone).
5) Above 2200 m mid-winter melt is negligible and many spring storms, which produce rain and melt at lower elevations, bring snow above 2200 m: one-third of the snow accumulation at 2340 m on the Tasman Glacier coincided with the ablation period at 1645 m in 1972 (Chinn, 1981) (spring accumulation/summer melt zone).

The elevations of these zones refer to the Waimakariri catchment, and decrease southward. For example, Fitzharris and Grimmond (1982) noted that the seasonal snowline in central Otago is at 1000 m. The elevations may also vary between years, in response to climatic variations. Snow accumulation patterns show a marked variation across the South Island, due



October 23, 1982 - 1200 GMT



October 31, 1982 - 1200 GMT

FIG. 7—Synoptic charts showing spring weather systems bringing (a) snowfall and (b) rain and melt. Upper chart is 7a; lower chart is 7b.

to the variability of snowfall during different types of storms. Data from the Waimakariri catchment are not available, but Chinn (1981) calculated an integrated index of snowpack magnitude and persistence, equal to the area under the water equivalent-time curve, for four snow course sites in the Waitaki catchment. The index for the site at 2340 m was poorly correlated with lower elevation sites, because of differences in timing of accumulation and melt, while the degree of correlation observed between pairs of the lower elevation sites could be explained largely by the relative exposures of the different sites to various storm types. The worst correlation was between the site with the greatest exposure to westerly storms and that with the greatest exposure to southerly storms.

The length and timing of the snow season and the magnitude of snow accumulation vary greatly from year to year (Fig. 6). Systematic variations in snowfall and accumulation on longer time scales (decades) have been mentioned by Burrows (1976), Tomlinson (1970), Hessel (1983) and La Chapelle (1979). However, Burrows and Tomlinson were discussing the concentrations of severe storms in the periods 1860 to 1880 and 1920 to 1940, rather than seasonal snow accumulation in the mountains, and definitive, long-term records of snow accumulation do not exist.

SNOWPACK INFLUENCES ON WATER MOVEMENT

O'Loughlin (1969) used dye to trace water movement through a snowpack at 1500 m in the Craigieburn Range; measured percolation rates ranged from 0.8 to 18 m/h. Moore (1984) similarly traced the movement of water through snow at Temple Basin at an elevation of 1500 m during a number of rain-on-snow and melt events during two snow seasons, and found that percolation rates were always greater than 1 m/h. Ice layers were relatively permeable: flow was diverted less than 3 m along sloping-impeding layers before flowing vertically into the underlying snow. Moore attributed this permeability to the frequent occurrence of melt and rain-on-snow near the main divide, which would promote ice layer decay (Gerdel, 1954). An experiment by Moore (1984) with snow tensiometers in ripe snow indicated a diurnal variation in capillary pressure between −70 and −130 mm H_20, which is similar to that recorded for mature mountain snow covers in western Canada (see for example Wankiewicz, 1978; Jordan, 1983).

Based upon the measured percolation rates and calculations based upon Colbeck's (1972) kinematic wave theory, Moore (1984) concluded that travel times for water movement through snow at Temple Basin should not exceed three hours for moderate rainfall intensities such as 10 mm/h, which is equivalent to the two-year return period twenty-four hour rainfall at Arthur's Pass (Tomlinson, 1980). Thus, percolation through snow should have minimal effect on the overall timing of water delivery during flood-producing events in the western areas of the Waimakariri catchment.

In eastern areas such as the Craigieburn Range, where smaller quantities of rain run through the snow, ice layers may be less permeable than observed at Temple Basin. Prowse and Owens (1984) observed a layer 80 mm thick in the Craigieburn Range, which was composed in some parts of clear ice; such layers may influence water movement during mid-winter rain-on-snow and melt events.

Water movement at the base of a snowpack is difficult to predict, but is partly controlled by the nature of the underlying surface and the presence of ground ice. Unfortunately, studies indicate that runoff processes cannot be uniquely related to surface characteristics (Taylor and Pearce, 1982; Mosley, 1982; Pierson, 1982; Griffiths and McSaveney, 1983). O'Loughlin (1964) described the occurrence of both concrete frost and superimposed ice on the eastern and western sides of the Craigieburn Range during a period without snow cover, while Moore (1984) and Burrows (1977) found that the ground under snow near the main divide was normally thawed and moist throughout winter.

Snow cover may have little effect on runoff routing in tussock areas, because the tussock supports the snow at least partially clear of the ground surface (Chinn, 1969; Weir, 1979), and would permit concentrated flow to occur in the gap between the base of the snow and the ground surface without interference. Such flow was observed at Temple Basin when snow pits were excavated during rain-on-snow events.

One final influence of snow on water movement is clogging of gorges by wind-drifted snow, which occurred during two field seasons at Temple Basin (Moore, 1984). In both years, the first major rain-on-snow event produced an outburst flood, similar in some ways to those described by Woo and Sauriol (1981) in the Arctic. These events produced greater deposition of sediment at the water level recorder at the outlet of the basin than any other type of event. Such snow-jam floods may have geomorphic significance in cirque basins in the Waimakariri catchment, but probably no more than local hydrological impact.

MONITORING AND MODELLING SNOW STORAGE AND MELT

Snowpack measurement

Both index and sample snow courses have limited utility in the Waimakariri catchment because of the great spatial variability of snow accumulation and melt, and the unpredictable timing of snowmelt events. Furthermore, the rugged terrain makes access to sampling points difficult and the avalanche hazard is often severe (Moore, 1984). Existing satellite images are unlikely to be useful because of insufficient spatial or temporal resolution to monitor variations in snow-covered areas and the frequency of cloud cover which decreases the number of usable images (Fitzharris, 1979). Methods such as snow pillows and nuclear snow gauges can supply the necessary temporal resolution, but lack spatial coverage, which is a limitation in a catchment where snowfall and snowmelt can occur simultaneously at different elevations.

Water balance methods

Changes in basin snow storage can in principle be calculated using water balance methods. For example, separation of snowmelt hydrographs has been a common tool for determining daily snowmelt totals. In the Waimakariri catchment, however, heavy rainfall often coincides with snowmelt, and the potentially large errors involved in assessing rainfall in exposed, mountainous watersheds contributes to the error in the estimate of snow storage change. Anderton (1974) tested a monthly water balance framework for operational estimates of snow storage in the upper Waitaki catchment. Although the results were promising, the techniques probably require some refinement. Water balance methods are best suited to estimating long-term mean monthly snow storage, as applied by Fitzharris and Grimmond (1982) and Bowden (1983). However, the use of mean monthly values masks year-to-year variability, which can be considerable and is important in management.

Modelling snow storage using climate data

Moore and Owens (1984b) applied a snow storage model to reproduce the snow survey data from Alan's Basin. Model performance was satisfactory when temperatures from Ski Basin (i.e. above the snowline) were extrapolated to the snow course elevation, but unsatisfactory when Craigieburn Forest temperatures (measured below the snowline) were used, because of marked and systematic deviations of actual lapse rate from the assumed rate of 6.5°C/km. The use of low elevation data to simulate higher elevation processes (as is common in many glaciological and snow hydrological studies) should be viewed circumspectly if assumed, constant lapse rates are used.

The results of Moore and Owens (1984b,c) support Fitzharris' (1979) suggestion that modelling snow storage using climate observations should prove fruitful. Using valley bottom precipitation data and temperature data measured at sites above and below the snowline (both hourly) would allow the calculation of at least a good index of snowpack storage for different elevation bands. Such an index could then be used in empirical relationships and streamflow analyses as suggested by Fitzharris (1979). If precipitation measurements from valley bottoms were extrapolated to mountain slopes by methods such as the "normal ratio method" as used by Kattelmann et al. (1985), these indices would more closely approximate the actual snow storages at different elevations, ignoring wind and avalanche redistribution. One advantage of using valley bottom precipitation data is that these sites are less prone to errors arising from low gauge catch efficiencies for snow.

When applying a snow storage model to large catchments like the Waimakariri, it may be necessary to use a horizontal distribution to account for the systematic variations in snow deposition and melt across the mountain ranges. This would contrast with previous applications of the HBV model to Scandinavian catchments, where a lumped model has been used to simulate snow storage over thousands of square kilometres (e.g. Bergstrom and Jonsson, 1976).

Parametric streamflow models

Although parametric streamflow models had been applied to South Island catchments in the 1960s and 1970s, none of them included a snow routine. Bowden (1974) suggested that unaccounted snow accumulation and melt were responsible for the overestimation of winter flows and underestimation of spring flows when Taylor's (1972) model was applied in the Waiau catchment. Fitzsimmons (1983) applied a version of Martinec's (1975) model to data from Camp Stream, using ground-based observations of snow cover. The modelling effort was not very successful, probably because of inadequate input data.

Moore and Owens (1984b) developed a model based on the Swedish HBV-3 model (Bergstrom and Jonsson, 1976), incorporating a snow routine distributed by elevation and vegetation type and a lumped transformation routine, and applied the model to the Camp Stream catchment using mean daily air temperature and precipitation as input. They found that mean daily air temperature was inadequate for discriminating rain from snow at elevations below 1500 m because of variations in precipitation phase during storms. Some aspects of the simulation results suggested that the transformation routine should be distributed by elevation or vegetation class, to reflect spatial variations in soil moisture and runoff mechanisms. A fundamental problem with the modelling exercise was that only twenty-three months of reliable data were available for the split-sample testing procedure, one year for calibration and eleven months for the independent test period. Ideally, both calibration and testing periods should include a number of years incorporating a range of hydrologic conditions.

SNOWMELT CONTRIBUTIONS TO SEASONAL STREAMFLOW

An estimate base upon basin hypsometry

Spring snowmelt may be estimated by combining estimates of the altitudinal variation of seasonal snow accumulation with catchment hypsometry through the following formula:

$$SP = \int_{1300\mathrm{m}}^{2400\mathrm{m}} P(z)\, F(z)\, (dA/dz)\, dz$$

where SP is the catchment mean seasonal snow accumulation (mm) at the end of winter, P(z) is the mean annual precipitation (mm) for a given elevation z (m), F(z) is the fraction of mean annual precipitation which is stored as snow at the end of winter at a given elevation, and A is the fraction of catchment area lying below an elevation z. The limits of integration are the lower limit of the lower seasonal snow zone and the highest point in the Waimakariri catchment.

The mean peak snow accumulation measured at Alan's Basin between 1962 and 1973 is approximately 500 mm, which represents twenty-eight percent of the mean annual precipitation in the Craigieburn Range. Anderton and Chinn (1978) found that maximum snow accumulation on the Ivory Glacier (elevation range 1400 to 1700 m; see Fig. 1) between 1969 and 1974 was on average twenty-five percent of annual precipitation. As a first approximation based on these values, F(z) is assumed to increase linearly from 0 at 1300m to 0.25 at 1500 m, then to remain constant with elevation in the upper portions of the basin, which experience less winter melt. Although precipitation varies greatly across the Southern Alps, the snow survey data cited do not contradict the assumption that F(z) is constant across the ranges when averaged over a number of years, although such variation may be important in a particular year.

If precipitation is assumed to be constant with elevation (this assumption will produce a conservative estimate of SP), then the fraction of annual precipitation which is stored as snow in the basin is given by the definite integral of F(z)(dA/dz). Using values of dA/dz from Fig. 2 yields a value of 0.06. Mean annual precipitation in the Waimakariri catchment is about 2300 mm, based upon measured discharges near the mouth, which were adjusted upward by fifteen percent to account for recharge losses to the Canterbury Plains and 500 mm for evapotranspiration (Bowden, 1983). Thus, approximately 140 mm of water is held as snow storage at the start of spring, and subsequently released as snowmelt, mainly during October and November of most years. This value of 140 mm represents thirty percent of the streamflow during October and November, which is ca. 460 mm (accounting for channel losses in the plains), or nine percent of mean annual streamflow.

Water balance estimates

Fitzharris (pers. comm.) has independently estimated peak spring snow accumulation based on a long-term mean monthly water balance approach using values of 2300 mm and 500 mm respectively for mean annual precipitation and evapotranspiration. He estimated that peak spring snow accumulation is on average 155 mm, which agrees roughly with the hypsometric estimate. The discrepancy between the two accords with the likely underestimation by the hypsometric method due to the assumption that precipitation does not vary with elevation.

The NCCBRWB (1986) also employed a long-term monthly water balance approach, using slightly different values for evapotranspiration and streamflow (600 and 1667 mm respectively). They estimated that seasonal snow storage represents fifteen percent of mean annual streamflow. The difference between this value and Fitzharris' probably results from differences in assumptions about changes in subsurface storage. The NCCBRWB calculations show snowmelt releases peaking in December and continuing through March, while field observations indicate that almost all of the seasonal snow cover disappears by late December in most years. This pattern may be an artifact of the way subsurface storage changes were calculated, because changes in snow storage were the residual in the water balance. The other possible explanation is that late-lying snow patches in topographic hollows and valley bottoms (such as the "Bealey glacier" avalanche deposit near Arthur's Pass) release meltwater throughout the summer. If this latter explanation is correct, then snowmelt would appear to be an important source of streamflow during low-flow periods in summer.

These figures suggest that the magnitude of snow storage is great enough that seasonal water yield models should account for snowmelt, and snowmelt contributions to streamflow in any given year and over the long term may be even greater than the derived mean values. Snow storage can be up to twice the mean value in some years (see Fig. 6), and if a reversal occurred of the current warming trend described by Burrows and Greenland (1979), a period of heavier snowfalls and a greater hydrologic significance for snow could result.

SNOWMELT CONTRIBUTIONS TO FLOODS

Snowmelt releases to a watershed depend on both snowmelt rates and snowcovered area. Insufficient data preclude an analysis of snowmelt volumes during floods in any drainage basin in the Waimakariri catchment. However, Fig. 8 demonstrates that the flow of 176 l/s at Camp Stream on 7 November 1982 (the highest daily flow that year) was generated purely by snowmelt over less than half the basin area; rain had not fallen for seven days, and foehn conditions were experienced in the Craigieburn Range that day (Moore, 1983). For comparison, the highest rainfall-generated daily flow in 1982 was 175 l/s on 19 November when over 50 mm of rain fell.

Fifty percent of the Waimakariri catchment and large portions of other eastern

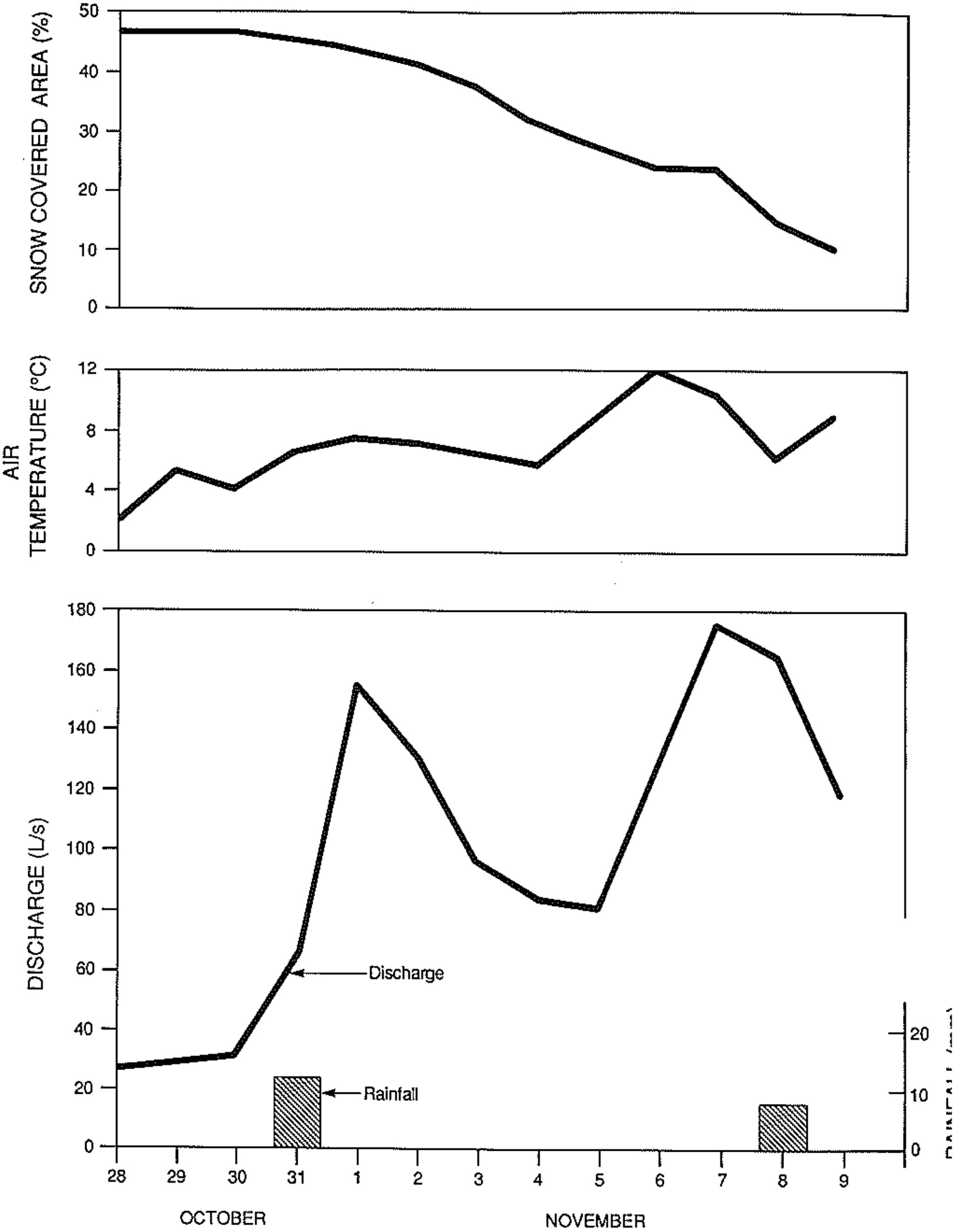


FIG. 8—Air temperature, rainfall, snow covered area and streamflow during spring snowmelt, Camp Stream.

South Island watersheds lie in the transient snow zone. Melting of transient snow covers during rain has been cited as an important cause of floods in other parts of the world (Harr, 1981; Cooley and Robertson, 1983; Johnson and Archer, 1973). Hydrological studies of the transient snow zone are difficult because of the often fleeting nature of important events. Meteorological and snowpack information with high temporal resolution (e.g. hourly data) are often not available.

Using chart records of wind speed, humidity and precipitation at Craigieburn Forest, Moore (1984) analysed one rain-on-snow event which resulted in flooding and road washouts in the Craigieburn area between 8 and 10 July 1983. Precipitation phases during the event were determined from six-hourly observations and the temperature record. Initial snowpack water equivalent was estimated from the observed snow depth preceding the event, using an assumed snow density of 300 kg/m^3 (a typical value for shallow snow which has been lying several days). Snowmelt was calculated from the U.S.A.C.E. (1956) equation for melt during heavy rain, which involves air temperature, wind speed and precipitation. The U.S.A.C.E. equation may underestimate melt in this environment (Moore, 1984), but is used here for lack of a more appropriate method.

Fig. 9 shows the input data and estimated snow storage and water releases to the ground surface for the event. Changes in snow storage have a marked effect on water release, significantly increasing it during periods of intense rain. The precipitation total for 9 July is 123 mm, while the estimated water release is 155 mm, an increase of over twenty-five percent. These figures can be compared to the twenty-four hour duration precipitation totals at Craigieburn Forest, which are 129 and 153 mm for return periods of five and ten years, respectively (NZMS, 1980). That is, the rainfall alone represents the input to be expected from a twenty-four hour rainfall at least once in five years, while the addition of snowmelt produces a water input which would be expected less than once in ten years by rainfall alone.

IMPLICATIONS FOR HYDROLOGICAL ANALYSES

Annual flood sequences in basins within the Waimakariri catchment can include floods generated by pure rainfall, rain-on-snow and pure snowmelt (e.g. the 7 November 1982 flow was a pure snowmelt flood, while the 8-10 July 1983 flows were generated be rain and snowmelt). Although flood sequences should be stratified be generating process before fitting probability distributions (Waylen and Woo, 1982; Waylen, 1982), insufficient data exist to permit such stratification of flood series for most South Island catchments.

The systematic spatial variability in snow storage and melt patterns implies that regional analyses of streamflow should consider variations in the hypsometry of the catchments and the distance from the main divide. For example, pure snowmelt and rain-on-snow floods may be relatively more important in the eastern portions of the catchment, where rainfalls are less intense, than in the western portions, where rainfall intensities likely far surpass the upper limits to snowmelt rates. Spatial variations in snowmelt contributions to flood hydrology may influence regionalised flood frequency curves.

Care should be taken to consider snow accumulation and melt in rainfall-runoff analyses. Changes in snow storage can alter the pattern of water releases

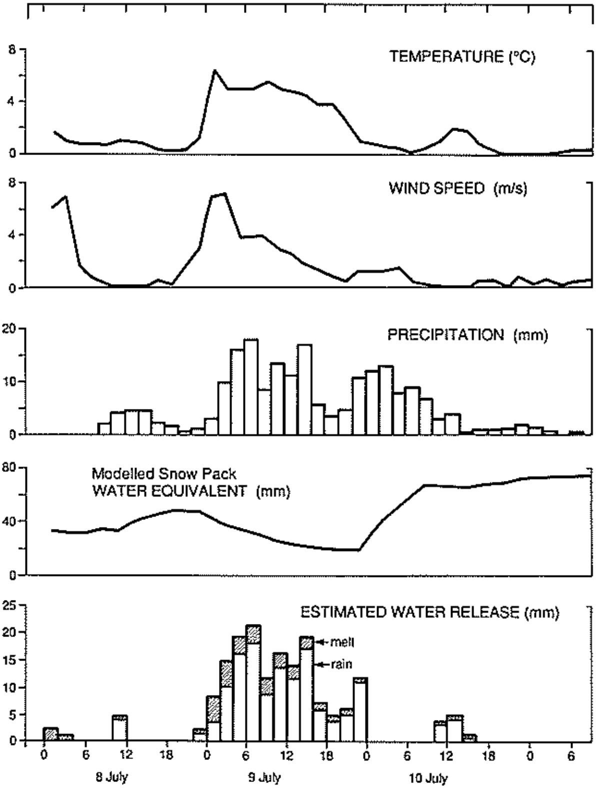


FIG. 9—Meteorological data and estimated snow storage and water release at Craigieburn Forest during a rain-on-snow event, 8 to 10 July 1983.

to a watershed, and could perturb the results of unit hydrograph derivations and runoff response analyses, such as those carried out in the Craigieburn area by Pearce and McKerchar (1979) and Taylor and Pearce (1982).

DIRECTIONS FOR FUTURE RESEARCH

In terms of practical significance, research should focus on two separate aspects. First is the assessment and prediction of the volume and timing of seasonal snowmelt. The accuracy (as opposed to precision) of the monthly snowmelt estimates of the NCCBRWB (1986) should be considered further. Snowpack indices calculated from climate data could be used in a purely correlational sense for streamflow prediction (e.g. Fitzharris, 1987), and as a component in operational water balance models. Work on water balance models which operate in near real time on daily to monthly intervals would allow a better understanding of yearly variations in seasonal snow and provide a tool for prediction of spring-summer streamflow.

The second focus should be assessment of the role of snowmelt in the flood hydrology of basins in the eastern portions of the catchment. The Craigieburn Range is an ideal site for such studies, having climate stations both above and below the seasonal snowline and existing gauging stations on Camp Stream and Hut Creek (which is adjacent to the Craigieburn Forest climate station and lies entirely within the transient snow zone). Additional observations including the snowline elevation and a recording snowmelt lysimeter at Craigieburn Forest would enhance research possibilities.

Studies should maximize the use of existing data, which have been collected at considerable cost and have not been exhausted in terms of research on snow hydrology (e.g. the streamflow data for the Bealey River near Arthur's Pass). Greater co-operation amongst government agencies, university researchers and others working in the mountains should be encouraged so as to increase the applicability of data to problems in snow hydrology. For example, data collected at ski fields for avalanche control purposes could be useful for snow hydrology research if made available in a standard format. Where possible, current data collections should be augmented to include relevant variables. For example, observers at climate stations within the mountains (e.g. Craigieburn Forest) could, at little additional cost, record the snowline elevation; this information would enable an estimate of snow-covered area, which is an important hydrological parameter.

ACKNOWLEDGEMENTS

Much of the work cited in this paper was carried out while the authors were doctoral candidates at the University of Canterbury, supported by Commonwealth Scholarships. Further financial assistance was received from the University Grants Committee and the Ministry of Works and Development for the National Water and Soil Conservation Organisation. The Forest Research Institute, the North Canterbury Catchment Board, the Ministry of Works and Development, the New Zealand Meteorological Service and the Department of Geography, University of Canterbury all provided logistical support and/or access to data. Mr C. S. L. Ommanney and Drs B. B. Fitzharris and P. Marsh made helpful suggestions on the manuscript. The authors are also grateful for the comprehensive reviews provided by Dr. M. J. McSaveney, T. J. H. Chinn, J. Barringer and an anonymous reviewer. Preparation of the manuscript was aided by funds from a Natural Sciences and Engineering Research Council operating grant to R. D. Moore.

REFERENCES

Anderton, P. W. 1973: The significance of perennial snow and ice to water resources of the South Island, New Zealand. *Journal of Hydrology (N.Z.) 12*: 6-18.

Anderton, P. W. 1974: Estimation of snow storage and melt in the catchment of Lake Pukaki. Paper presented at the New Zealand Hydrological Society Annual Symposium, Dunedin.

Anderton, P. W.; Chinn, T. J. 1978: Ivory Glacier, New Zealand, an I.H.D. representative basin study. *Journal of Glaciology 20:* 67-84.

Bergstrom, S.; Jonsson, S. 1976: The application of the HBV runoff model to the Filefjell research basin. Rapporter Nr RHO 5, Swedish Meteorological and Hydrological Institute, Norrkoping.

Bowden, M. J. 1974: The water resources of the Waiau catchment. North Canterbury Catchment Board, Christchurch.

Bowden, M. J. 1977: The water resources of the Hurunui catchment. North Canterbury Catchment and Regional Water Board, Christchurch.

Bowden, M. J. 1983: Ice and snowfields. In: *Rakaia Resource Report, Vol. 2.* North Canterbury Catchment and Regional Water Board, Christchurch: 7-10.

Burrows, C. J. 1976: Exceptional snow storms in the South Island High Country. *Tussock Grasslands and Mountain Lands Institute Review 32*: 43-47.

Burrows, C. J. 1977: Alpine grasslands and snow in the Arthur's Pass and Lewis Pass regions, South Island, New Zealand. *New Zealand Journal of Botany 15*: 665-686.

Burrows, C. J.; Greenland, D. E. 1979: An analysis of the evidence for climatic change in New Zealand in the last thousand years: evidence from diverse natural phenomena and from instrumental records. *Journal of the Royal Society of New Zealand 9*: 321-373.

Chinn, T. J. H. 1969: Snow survey techniques in the Waitaki catchment, South Canterbury. *Journal of Hydrology (N.Z.) 8*: 68-76.

Chinn, T. J. 1987: Snowfall variations, hazards and snowmelt. Paper presented at the Mountain Lands Workshop, Lincoln College, Lincoln, New Zealand.

Chinn, T. J., Whitehouse, I. E. 1980: Glacier snowline variations in the Southern Alps, New Zealand. In: World Glacier Inventory, *IAHS-AISH Publication No. 126*, Washington: 219-228.

Colbeck. S. C. 1972: A theory of water percolation in snow. *Journal of Glaciology 11*: 369-385.

Cooley, K. R.; Robertson, D. C. 1983: Monitoring a rain-on-snow event. Proceedings of the 51st Western Snow Conference, Vancouver, Washington: 19-26.

Dalmer, E. 1971: The Waimakariri as a water resource. North Canterbury Catchment Board, Christchurch.

Fitzharris, B. B. 1978: Problems in estimating snow accumulation with elevation on New Zealand mountains. *Journal of Hydrology (N.Z.) 17*: 78-90.

Fitzharris, B. B. 1979: Snow hydrology. In: Murray, D. L. and Ackroyd, P. (eds.) *Physical Hydrology — New Zealand Experience*, New Zealand Hydrological Society, Christchurch: 23-43.

Fitzharris, B. B. 1987: A method for indexing the variability of alpine seasonal snow over large areas. In: Large-scale effects of seasonal snow cover, *IAHS-AISH Publication No. 166*, Washington: 139-150.

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, *IAHS-AISH Publication No. 138*, Washington, p. 161-168.

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

Fitzsimmons, S. 1983: Modelling snowmelt runoff in Camp Stream catchment. Unpublished B. Sc. (Hons) project, Univ. Canterbury, Christchurch.

Gerdel, R. W., 1954: The transmission of water through snow. *Transactions of the American Geophysical Union 35*: 475-485.

Griffiths, G. A.; McSaveney, M. J. 1983: Hydrology of a basin with extreme rainfalls — Cropp River, New Zealand. *New Zealand Journal of Science 26*: 293-306.

Harding, F. B. 1972: Micro-meteorological investigations over a mid-latitude temperate glacier — the Ivory Glacier. Unpublished M.A. thesis, Univ. Canterbury, Christchurch.

Harr, R. D. 1981: Some characteristics and consequences of snowmelt during rainfall in Western Oregon. *Journal of Hydrology 53*: 277-304.

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

Harrison, W. 1986b: Effects of snow fences on the snowpack of a block mountain in Otago. *Journal of Hydrology (N.Z.) 25*: 18-40.

Hayward, J.A. 1967: The Waimakariri catchment. *Special Publication No. 5*, Tussock Grasslands and Mountain Lands Institute.

Hessel, J. W. D. 1983: Climatic effects on the recession of the Franz Josef Glacier. *New Zealand Journal of Science 26*: 315-320.

Johnson, P.; Archer, D. R. 1973: The significance of snow in Britain. In: The role of snow and ice in hydrology, *IASH Publication No. 107, Vol. 2*, Paris: 1098-1110.

Jordan, R. P. 1983: Meltwater movement in a deep snowpack, 1, field observations. *Water Resources Research 19*: 971-978.

Kattelmann, R. C.; Berg, N. H.; Pack, M. K. 1985: Estimating regional snow water equivalent with a simple simulation model. *Water Resources Bulletin 21*: 273-280.

LaChapelle, E. R. 1979: An assessment of avalanche problems in New Zealand. *Avalanche Committee Report No. 2,* New Zealand Mountain Safety Council, Wellington.

Lyall, I. T. 1972: The Polar Low over Britain. *Weather 27*: 378-390.

Marcus, M. G.; Moore, R. D.; Owens, I. F. 1985: Short term estimates of surface energy transfers and ablation on the lower Franz Josef Glacier, South Westland, New Zealand. *New Zealand Journal of Geology and Geophysics 28*: 559-567.

Martinec, J. 1975: Snowmelt-runoff model for streamflow forecasts. *Nordic Hydrology 6*: 145-153.

McGregor, G. R. 1984: Snow avalanche phenomena in the Craigieburn Range, New Zealand. Unpublished PhD. thesis, Univ. Canterbury, Christchurch, New Zealand.

McSaveney, M. J. 1978: The magnitude of erosion across the Southern Alps. In: *Erosion assessment and control in New Zealand,* Proceedings of a conference at Christchurch: 7-35.

Moore, R. D. 1983: A comparison of the snowmelt energy budgets in two alpine basins. *Archives for Meteorology, Geophysics and Bioclimatology Series A 33:* 1-10.

Moore, R. D. 1984: Snow hydrology of a mountainous rain-on-snow environment — the Waimakariri catchment, New Zealand. Unpublished Ph.D. thesis, University of Canterbury, Christchurch, New Zealand.

Moore, R. D.; Owens, I. F. 1984a: Controls on advective snowmelt in a maritime alpine basin. *Journal of Climate and Applied Meteorology 23*: 135-142.

Moore, R. D.; Owens, I. F. 1984b: Modelling alpine snow accumulation and ablation using daily climate data. *Journal of Hydrology (N.Z) 23*: 73-83.

Moore, R. D.; Owens, I. F. 1984c: A conceptual runoff model for a mountainous rain-on-snow environment, Craigieburn Range, New Zealand. *Journal of Hydrology (N.Z.) 23*: 84-99.

Morris, J. Y.; O'Loughlin. C. L. 1965: Snow investigations in the Craigieburn Range. *Journal of Hydrology (N.Z.) 4*: 2-16.

Mosley, M. P. 1982: Subsurface flow velocities through selected forest soils, South Island New Zealand. *Journal of Hydrology 55*: 65-92.

Neale: A. A.; Thompson, G. H. 1977: Meteorological conditions accompanying heavy snowfalls in Southern New Zealand. *Technical Information Circular No. 155*, New Zealand Meteorological Service, Wellington.

NZMS (N.Z. Meteorological Service) 1980: Depth-duration-frequency tables based on daily rainfalls. *Miscellaneous Publication No. 162 Supplement 1*, New Zealand Meteorological Service, Wellington.

N.Z. National Committee for the I.H.D. 1969: Preparatory report of the Technical Subcommittee on Snow. Wellington, New Zealand.

NCCBRWB (North Canterbury Catchment Board and Regional Water Board) 1986: Waimakariri River and Catchment Resource Survey I. NCCBRWB, Christchurch.

O'Loughlin, C. L. 1964: Notes on the occurrence of concrete frost and other forms of soil-ice in Canterbury beech forests. *Protection Forestry Report No. 5*, New Zealand Forest Service, Forest and Range Experiment Station, Rangiora (unpublished).

O'Loughlin, C. L. 1969: Further snow investigations in the Craigieburn Range. *Protection Forestry Report No. 52*, New Zealand Forest and Range Experiment Station, Rangiora (unpublished).

Owens, I. F.; McGregor, G.; Prowse, T. D. 1983: Snow avalanche hazards in the Canterbury high country. In: Bedford, R. D. and Sturman, A. P. (eds.) *Canterbury at the Crossroads*, Miscellaneous Series No. 8, New Zealand Geographical Society, Christchurch: 166-181.

Pearce, A. J.; McKerchar, A. I. 1979: Upstream generation of storm runoff. In: Murray, D. L. and Ackroyd, P. (eds.) *Physical Hydrology — New Zealand Experience*, New Zealand Hydrological Society, Christchurch: 165-193.

Pierson,. T. C. 1982: Classification and hydrological characteristics of scree slope deposits in the northern Craigieburn Range, New Zealand. *Journal of Hydrology (N.Z.) 21*: 34-60.

Prowse, T. D. 1981: The snow environment of the Craigieburn Range. Unpublished Ph.D. thesis, Univ. Canterbury, Christchurch, New Zealand.

Prowse, T. D.; Owens, I. F. 1982: Energy balance over melting snow, Craigieburn Range, New Zealand. *Journal of Hydrology (N.Z.) 21*: 133-147.

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.

Rowe, L. K. 1968: Summary of surface wind data, Craigieburn Range 1961-1967. *Protection Forestry Report No. 45*, New Zealand Forest Research Institute, Christchurch, New Zealand.

Taylor, C. H.; Pearce, A. J. 1982: Storm runoff processes and subcatchment characteristics in a New Zealand hill country catchment. *Earth Surface Processes and Landforms 7*: 439-447.

Taylor, M. A. 1972: Assessment of a mathematical model for runoff prediction in New Zealand. Unpublished M. E. thesis, University of Canterbury, Christchurch, New Zealand.

Tomlinson, A. I. 1970: The Canterbury snowfall of November 1967. *New Zealand Geographer 26*: 20-35.

Tomlinson, A. I. 1980: The frequency of high intensity rainfalls in New Zealand, Part I. *Water and Soil Technical Publication No. 19*, National Water and Soil Conservation Organisation, Wellington, New Zealand.

U.S.A.C.E. (U.S. Army Corps of Engineers) 1956: Snow Hydrology. North Pacific Division, Portland, Oregon.

Wankiewicz, A. 1978: Water pressure in ripe snowpacks. *Water Resources Research 14*: 593-599.

Waylen, P. 1982: Some characteristics and consequences of snowmelt during rainfall in western Oregon—a comment. *Journal of Hyrdology 58*: 185-188.

Waylen, P.; Woo, M. -K. 1982: Prediction of annual floods generated by mixed processes. *Water Resources Research 18*: 1283-1286.

Weir, P. L. 1979: Topographic influences on snow accumulation at Mount Hutt. Unpublished M.Sc. thesis, University of Canterbury, Christchurch, New Zealand.

Woo, M. -K.; Sauriol, J. 1981: Effects of snow jams on fluvial activities in the high Arctic. *Physical Geography 2*: 83-98.

Younkin, R. J. 1968: Circulation patterns associated with heavy snowfall over the Western United States. *Monthly Weather Review 96*: 851-853.