The contribution of snowmelt to the rivers of the South Island, New Zealand

Tim Kerr
National Institute of Water and Atmospheric Research (NIWA), PO Box 8602, Christchurch. Corresponding author: timkerr37@hotmail.com

Abstract
A new assessment of the mean annual (1981–2000) stream flow derived from snow and ice melt ($Q_m$) as a proportion of mean annual stream flows ($Q_m/Q$) for the South Island of New Zealand has been prepared. The assessment allows identification of the meltwater contribution of flows for any reach in the country, including bridge, irrigation take and lake locations. The snow melt estimates were generated by separating gridded daily precipitation into rainfall and snowfall components to provide mean annual snowfall and rainfall data. The ratio of the catchment-based accumulation of these precipitation components was then found, which is shown to be equivalent to the ratio of melt-derived flow to total flow. The assessment allows comparisons between sites, which were previously complicated by a paucity of assessments and differing methods. Based on the assessment, 3% of the South Island's mean annual stream flow is made up of meltwater. This is considerably less than estimates for regions of California, Austria, India and Chile. This difference is probably a reflection of the relative lack of seasonality in precipitation and the generally mild winter temperatures in the South Island. This low snow and ice melt contribution is problematic for irrigation schemes, in that artificial storage management is required if winter precipitation is to be retained for summer plant growth, but beneficial for hydro-electricity generation in that stream flows continue throughout the year, including winter (albeit often reduced) when energy demands are often high.

The melt percentage data is available as an Excel™ spreadsheet via http://dc.niwa.co.nz/niwa_dc/srv/en/metadata.show?uuid=02f3e247-2f78-562a-8712-675b4f3a0166.

Keywords
Snow, melt, stream flow, South Island, alpine, precipitation

Introduction
A stream’s flow regime is affected by the relative contribution of snowmelt (Dettinger and Diaz, 2000). Snowmelt affects flow seasonality and diurnal variability, and influences flood flows. In New Zealand's South Island, snow melt enhances the base flow of many rivers during the summer irrigation season, but snow storage suppresses flow in many hydro-electric catchments during winter when generation demand is high (Gillies, 1964). Flood magnitudes are enhanced when snowmelt augments rainfall events (Fitzharris et al., 1980; Moore and Prowse, 1988) but are suppressed in winter in those catchments where precipitation falls as snow and so goes into snow storage (Jowett and Thompson, 1977). In addition, if there is a large cold snowpack, rain may be absorbed and frozen into the snow pack (Colle et al., 2013). To anticipate likely flood magnitudes
and frequency and to properly understand a stream’s flow regime, it is necessary to consider the likely contribution of snowmelt to a stream’s flow. This is reflected in the New Zealand guidelines for bridge building, which require special consideration to be taken when estimating possible floods in rivers that have a snow-melt contribution (Transit New Zealand, 2003).

Where there is a strong winter input to precipitation seasonality, snowmelt contribution to stream flow may be estimated from the seasonal hydrograph. For example in Californian mountain catchments, with most precipitation falling as snow in the wintertime, snowmelt may be considered to be the April–July (spring/summer) component of the annual hydrograph (Roos, 1991). In New Zealand, however, spring flows may originate from either snow melt or spring rainfall, so hydrograph separation based on season is not possible. This difficulty has led to the frequent use of the water balance to determine the contribution of snowmelt to stream flow in New Zealand catchments (e.g., Anderton, 1974; Jowett and Thompson, 1977; Fitzharris and Grimmond, 1982; Bowden et al. 1983; Cowie et al., 1986; McKerchar et al., 1998).

In the South Island of New Zealand snow fall is more common in the mountains than in the lowlands. The frequency of days with snow on the ground is shown in Figure 1, which is derived from the 8-day composite satellite-image-based MOD10A2.5 snow cover for the 2000–2011 period (Hall et al., 2006, updated weekly), with cloud removal processing by Pascal Sirguey, University of Otago, following Dozier et al. (2008). The importance of snow to hydrology is likely to be greater than the snow cover frequency suggests because the island’s precipitation distribution is controlled by orographic effects, leading to a strong west-to-east gradient (Fig. 1). Rivers flowing to the east from the Southern Alps may have the larger part of their catchment

---

**Figure 1** – Elevation, mean annual precipitation (1951–1980) and frequency of snow cover (2000–2011) of the South Island.
in areas where little snow falls, but the source of most of their water is within their western mountainous sub-catchments where snow is more frequent.

Previous assessments of the contribution of meltwater to stream flows in catchments of the South Island (Fig. 2, Table 1) range from 33% in the small alpine catchment of the Fraser (Fitzharris and Grimmond, 1982) to 9% for the large catchment of the Waimakariri (Moore and Prowse, 1988). The difference between these assessments may be attributed to their relative catchment areas that are outside of common snowfall zones. In contrast, the difference between 33% for the Fraser and 21% for the Ivory Glacier basin (Anderton and Chinn, 1978) is notable because both catchments are alpine in nature. In this case the difference is likely to be related to their differing climate, the glacial recession in the Ivory, and the different parameters reported. The Ivory is located in the high precipitation, light wind, mild temperatures of Westland (Hessel, 1982), whereas the Fraser is in a low precipitation, strongly seasonal temperature region of Central Otago (de Lisle and Browne, 1971). This would indicate that the proportion of precipitation that falls in winter is more likely to fall as snow in the Fraser than in the Ivory, even though the overall precipitation magnitudes are much larger in the Ivory. This would tend to make meltwater a greater component of stream water in the Fraser. Glacial recession within the Ivory accounted for 9% of its meltwater (Anderton and Chinn, 1978), whereas the Fraser has no permanent snow or ice fields. This indicates that under equilibrium conditions the meltwater portion of the streamflow in the Ivory is just 12%, an even greater difference to the Fraser estimate. The third explanation for the difference between the Fraser and the Ivory is that the Fraser’s number is an estimate of the snow melt, $M$, as a portion of stream flow, $M/Q$, and the Ivory’s number is a measure of the meltwater in the stream $Q_m$ as a portion of the stream flow, $Q_m/Q$. The Fraser’s value does not account for evapotranspiration of the snowmelt before it reaches the stream, which would lower the Fraser estimate closer to the value estimated for the Ivory.

While the previous assessments show a diversity of importance of snowmelt to catchment hydrology, they provide little indication of how important snowmelt is to the remainder of the country or for sub-catchments of those areas that have been assessed. In addition, the different parameters reported, and the different methods used makes comparison difficult.

**Table 1** – Seasonal temperature lapse rates used in interpolating station data to the regular Virtual Climate Station Network (VCSN) grid to produce the Norton temperatures

<table>
<thead>
<tr>
<th></th>
<th>Maximum temperature ($^\circ C m^{-1}$)</th>
<th>Minimum temperature ($^\circ C m^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer (DJF)</td>
<td>0.0061</td>
<td>0.0041</td>
</tr>
<tr>
<td>Autumn (MAM)</td>
<td>0.0063</td>
<td>0.0032</td>
</tr>
<tr>
<td>Winter (JJA)</td>
<td>0.0064</td>
<td>0.003</td>
</tr>
<tr>
<td>Spring (SON)</td>
<td>0.0066</td>
<td>0.0042</td>
</tr>
</tbody>
</table>

Figure 2 – Catchments in which previous assessments of snowmelt have been made.
The recent availability of national-scale precipitation and temperature grids (Tait et al., 2006; Tait, 2008), in concert with advances in computing speed and storage that have simplified hydrological accumulation modelling, enables a national-scale assessment of the relative importance of meltwater to stream flow. This means that the estimation of snow-melt contribution to mean annual flows is no longer limited to those catchments that have had intensive hydrological investigations, but may be provided for any stream reach in the country in a consistent manner.

This paper presents a national assessment of meltwater contribution to mean annual stream flows for New Zealand, and the methods and data used in preparing the assessment. The uncertainty of the results is discussed and compared to previous assessments and to assessments from other regions of the world.

Methods

The water balance for a catchment may be described by equating precipitation, \( P \), with evapotranspiration, \( ET \), stream flow, \( Q \), and the change in water storage, \( \Delta S \) (Toebes, 1972):

\[
P = ET + Q + \Delta S
\]  

(1)

Separating precipitation into rainfall, \( P_r \), and snowfall, \( P_s \), and dividing the water storage into snow and ice storage, \( \Delta S_s \), and all other storage, \( \Delta S_o \) (e.g., ground water, soil water, and canopy water) and solving for flow leads to:

\[
Q = P_r + P_s + \Delta S_s + \Delta S_o - ET
\]  

(2)

The change in snow storage, \( \Delta S_s \), equals the snowfall, \( P_s \), less the snow melt, \( M \):

\[
\Delta S_s = P_s - M
\]  

(3)

The water balance components are often calculated at a monthly time scale, estimating evapotranspiration from temperature and using various estimates for water storage (e.g., Fitzharris and Grimmond, 1982; Bowden et al., 1983; Cowie et al., 1986).

At an annual time step, under constant land cover and climate, the change in snow and ice, and other storage may be considered to average out to zero, so that annual flow may be derived from annual evapotranspiration and precipitation

\[
Q = P_r + P_s - ET
\]  

(4)

Evapotranspiration may be divided into constituent parts originating from rainfall, \( ET_r \), and snow and ice melt, \( ET_s \):

\[
ET = ET_s + ET_r
\]  

(5)

The meltwater contribution to a stream, \( Q_m \), may be considered to be derived from the snow and ice melt, \( M \), less the evapotranspiration that originates from the melted snow and ice, \( ET_s \):

\[
Q_m = M - ET_s
\]  

(6)

Combining this with equation (3) and setting the annual change in snow storage to zero so that snowmelt, \( M \), equals snow accumulation, \( P_s \), leads to:

\[
Q_m = P_s - ET_s
\]  

(7)

The ratio of stream flow from melt relative to total flow may be obtained from equations (7) and (4):

\[
\frac{Q_m}{Q} = \frac{P_s - ET_s}{P - ET}
\]  

(8)

Under the assumption that \( ET_r \) and \( ET_s \) are in proportions equal to the relative contributions of rainfall and snowfall, that is:

\[
ET_s = ET \frac{P_s}{P}
\]  

(9)

Then by substituting \( ET_s \) using equation (9) into equation (8) and simplifying, the ratio
of flow from meltwater to total flow becomes:

$$\frac{Q_m}{Q} = \frac{P_s}{P}$$  \hspace{1cm} (10)

This enables the contribution of melt to stream flows to be determined through estimates of the catchment's total frozen precipitation, $P_s$, and total precipitation, $P$.

Daily 5-km gridded precipitation data for New Zealand is available from 1960 from the Virtual Climate Station Network (VCSN) (Tait et al., 2006) provided by Andrew Tait, NIWA Wellington. These data are prepared by interpolating observed daily rainfall using a thin plate smoothing spline and guided by the 1951–1980 mean annual precipitation distribution (New Zealand Meteorological Service, 1985). The snowfall component of these precipitation data were taken as being at the locations and on the days when the daily temperature was below 0°C. The daily temperature was defined as the average of the daily maximum and minimum temperature. The daily maximum and minimum temperature were obtained from the ‘Norton’ VCSN maximum and minimum temperatures. The ‘Norton’ VCSN temperature data were prepared by converting available daily climate station maximum and minimum temperatures to sea level values using a seasonally varying lapse rate, interpolating these values to the VCSN’s 5-km grid using a thin plate smoothing spline, then adjusting the temperature at each grid point to the average elevation of the VCSN grid square, using the same lapse rate. The lapse rates originate from the work of Norton (1985) and are given in Table 1

Snowfall and total precipitation values for each day and grid point were used to provide 1981–2010 mean annual totals, which were then resampled to a 30-m grid using bilinear interpolation.

To determine $Q_m/Q$, it is necessary to determine the precipitation that accumulates in every stream reach from the surrounding terrain and the upstream reaches. This was achieved by generating 30-m resolution ‘flow accumulation’ grids for New Zealand through GIS processing using ArcGIS™. A terrain flow path grid, required to calculate the flow accumulations, was generated using a 30-m digital elevation model (DEM) of the country that was originally generated from the 20 m contours of the NZ topographical data (LINZ, 2000). To ensure the flow paths were hydrologically correct, and not limited by the resolution of the DEM (in which narrow gorges may not be resolved), the mapped stream network for New Zealand, as provided in the River Environment Classification (REC) (Snelder et al., 2004), was ‘burnt’ into the DEM using the methods of Saunders (1999). This hydrologically correct flow direction grid enabled the generation of flow accumulation grids of the snowfall and total precipitation. The flow accumulation for the snowfall was divided by the flow accumulation for the total precipitation, providing an estimate of $Q_m/Q$ as per equation (10).

**Results**

An Excel™ file of Qm/Q percentages for each River Environment Classification New Zealand reach ID is available via http://dc.niwa.co.nz/niwa_dc/srv/en/metadata.show?uuid=02f3e247-2f78-562a-8712-675b4f3a0166. A map of $Q_m/Q$ for South Island rivers is shown in Figure 3. For clarity, only river reaches with more than three upstream branches on the underlying River Environment Classification (i.e., strahler 4) are shown. The map shows that a large part of the island’s streams have less than 0.5% snowmelt contribution (rounded down to zero). These are all the ‘low country’ streams. The streams with higher snowmelt contributions are either in,
or originate from elevated country. Overall, 3.3% of the water that flows from the South Island’s streams into the sea comes from snow-melt. A selection of larger rivers from each region of the South Island that flow to the sea is shown in Figure 3. The Waitaki has the highest meltwater portion of streamflow with 17%, followed by the Rangitata with 15%. The Waitaki is a nationally important river for hydro-electricity generation, so the high contribution of snowmelt is problematic in that the resulting low flows in winter coincide with a period of high electricity demand. In contrast, the Rangitata is a regionally significant source of irrigation water, so the high snowmelt is beneficial for flow timing, in that the winter snow acts as a reservoir for the irrigation systems. The high snow melt for the Clarence (8%) demonstrates that not all of the large snowmelt rivers must come from the higher mountains of the Southern Alps.

The $Q_m/Q$ at river reaches associated with a range of highway bridges is shown in Figure 4. Many of the major highways are located near the coast, so the snowmelt contribution is close to the minimum river mouth values. The higher snowmelt values

![Figure 3](image-url) – Snowmelt contribution to mean annual flows for rivers of strahler order 4 and higher. Percentages given in brackets are the estimated snowmelt contribution at the river mouths for a selection of larger rivers.
for state highway bridges are either on inland roads that run beside high mountain ranges, for example the bridge over Wye Creek on State Highway 63 (24%), or have a nearby glacier source, e.g., the Fox (24%) and Waiho (19%) rivers. As set out in the design requirements for bridges in New Zealand, it is necessary to know if a stream reach associated with a bridge is influenced by snow melt to know which method of flood flow estimation needs to be applied. When there is a contribution from snow melt, then a detailed hydrological investigation is required, rather than the standard application of the rational or regional method (Transit New Zealand, 2003).

Snowmelt contribution at the river reaches associated with some of the larger irrigation intakes are shown in Figure 5. This figure identifies those schemes that benefit from the winter water storage in the snow pack. While it has previously been reported that the Fraser Dam has a significant snow melt component, it is probably not as commonly known that the Opuha Dam, which is located in a relatively low-lying region (390 m) has a seventh of its water originating from snow melt. This enables the scheme to operate much more

Figure 4 – Contribution of snowmelt to stream flows at a selection of river reaches associated with state highway (SH) bridges. Black lines represent the highway road network.
efficiently than if all of its water came from rain alone. While the Waitaki’s Bortons Pond intake has a large snowmelt component, this intake is downstream of several river storage and control structures, so the seasonal flow effects experienced there are more to do with the upstream reservoir management than with the effects of melt. In contrast, the Waiau, Ashburton and Rangitata intakes have no storage and so are highly reliant on the winter snow storage to help augment base flows during the spring growth. To some degree, all of these schemes need to consider improvements in efficiency, reduced supply, or increased reservoir capacity to allow for the potential reduction in winter snow storage under a warming climate.

The snowmelt contribution to some of the larger lakes in the South Island is shown in Figure 6. This clearly identifies the three hydro-lakes of the Waitaki as having a significant snowmelt contribution. For Lakes Pukaki and Tekapo, the related seasonal snow storage is managed to some degree by their large controllable storage. Most of the other lakes have a relatively small controllable storage, primarily for flood control, so that seasonal water storage is dominated by the snow storage. Lake Rotoroa in particular is of interest, as it has no storage control and a
relatively high snow melt contribution (10%). This means that the Gowan, which drains Lake Rotoroa, is one of the few natural-flow rivers in the country that drains a large lake with a snow melt source.

Discussions

Comparison between the results of this study and previous assessments are provided in Table 2. This shows that the new estimations are a reasonable approximation to the previous estimates. Care must be taken when comparing results to ensure that the same measure is being used. As outlined earlier, the reported figure of 33% for the Fraser snowmelt (Fitzharris and Grimmond, 1982) is for the maximum accumulated snow as a proportion of total annual stream flow. This is different from $Q_m/Q$ reported here, which is an assessment of the snow melt that gets to the stream. The difference between the two is the evapotranspiration of the melted snow between the snow pack and the stream channel, as outlined in Equation (6). If equation (10) is applied to the mean maximum annual snow accumulation (176 mm) and precipitation (869 mm) estimates from Fitzharris and Grimmond (1982), then their $Q_m/Q$ is 20%, similar to the estimates from this national assessment.

Figure 6 – Snow melt contribution to mean annual inflows at a selection of South Island lakes
Table 2 – Contribution of snow and ice melt to catchment flow from previous estimates and this work. \( M/Q \) is mean annual snow melt as a portion of mean annual stream flow. \( Q_m/Q \) is the percentage stream flow that is meltwater, based on estimates of mean annual snow accumulation, \( P_s \), and mean annual precipitation, \( P \), following equation (10). Locations are shown in Figure 2.

<table>
<thead>
<tr>
<th>Area</th>
<th>( M/Q ) (%)</th>
<th>( Q_m/Q ) (%)</th>
<th>Citation</th>
<th>( Q_m/Q ) (%) This study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fraser Basin</td>
<td>33</td>
<td>20</td>
<td>(Fitzharris and Grimmond 1982)</td>
<td>22</td>
</tr>
<tr>
<td>Lake Pukaki</td>
<td>19</td>
<td></td>
<td>(Anderton 1974)</td>
<td>23</td>
</tr>
<tr>
<td>Combination of Lakes Manapouri, Te Anau, Waipori, Hawea, Benmore, Pukaki, Tekapo, Cobb and Coleridge</td>
<td>21</td>
<td>17</td>
<td>(Fitzharris 1987)</td>
<td>14</td>
</tr>
<tr>
<td>Lakes Tekapo, Pukaki and Ohau</td>
<td>22</td>
<td>17</td>
<td>(McKerchar et al., 1998)</td>
<td>21</td>
</tr>
<tr>
<td>Lakes Wanaka, Hawea and Wakatipu</td>
<td>16</td>
<td>11</td>
<td>(McKerchar et al., 1998)</td>
<td>12</td>
</tr>
<tr>
<td>Lakes Wanaka, Hawea and Wakatipu</td>
<td>14</td>
<td></td>
<td>(Jowett and Thompson, 1977)</td>
<td>12</td>
</tr>
<tr>
<td>Clutha at Roxburgh</td>
<td>14</td>
<td>11</td>
<td>(McKerchar, 1997)</td>
<td>12</td>
</tr>
<tr>
<td>Clutha at Roxburgh</td>
<td>14</td>
<td>10</td>
<td>(Jowett and Thompson, 1977)</td>
<td>12</td>
</tr>
<tr>
<td>Clutha at Balclutha</td>
<td>11</td>
<td></td>
<td>(Poyck et al., 2011)</td>
<td>10</td>
</tr>
<tr>
<td>Waimakariri at Gorge</td>
<td>15</td>
<td>11</td>
<td>(Cowie et al., 1986)</td>
<td>9</td>
</tr>
<tr>
<td>Waimakariri at Old Highway Bridge</td>
<td>9</td>
<td>6</td>
<td>(Moore and Prowse, 1988)</td>
<td>8</td>
</tr>
<tr>
<td>Ivory Glacier</td>
<td>21</td>
<td></td>
<td>(Anderton and Chinn, 1978)</td>
<td>15</td>
</tr>
<tr>
<td>Rakaia at Gorge</td>
<td>15</td>
<td>11</td>
<td>(Bowden et al., 1983)</td>
<td>12</td>
</tr>
<tr>
<td>Rakaia</td>
<td>11</td>
<td></td>
<td>(McSaveney, 1984)</td>
<td>11</td>
</tr>
</tbody>
</table>

of 22%. Likewise, the water balance figures for the Rakaia River (Bowden et al., 1983) indicate maximum snow storage is the equivalent to 14% of the mean annual flows, but that \( Q_m/Q \) (according to equation (10)) is just 11%, which is well approximated by the 12% estimated here. For the Waimakariri, the reported 15% of annual flows (Cowie et al., 1986) is again based on the accumulated snow storage, and not the meltwater portion of the river’s flow. Applying equation (10) to their precipitation and snow accumulation estimates returns a \( Q_m/Q \) of 11%, which is near the 9% estimated here. Where previous estimates have been of \( M/Q \), the published values of mean annual snow accumulation and mean annual precipitation have been used to calculate \( Q_m/Q \) using equation (10). The difficulty with comparing these various estimates is that they all use slightly different methodologies. For instance, a crucial part of applying a monthly water balance is the estimation of the water storage components separate from snow storage. Fitzharris and Grimmond (1982) use their expert knowledge to allocate a catchment storage component separate to snow as being up to a maximum of 75 mm across their catchment and it is full when snow storage is at a maximum. In contrast, the method of Cowie et al.
assessed non-snow catchment storage using a master recession flow technique that assumes that flows of the same magnitude in recession have the same non-snow storage. Yet another approach to catchment storage is taken by Bowden et al. (1983) whereby all catchment storage is considered to be snow storage. Through using a single system for the entire country, relative comparisons between catchments may be undertaken without correction for subtleties of methodology.

Equation (10) provides an estimate for when snow and ice storage is on average in a state of equilibrium. New Zealand’s glaciers, however, are in a general state of recession (Chinn et al., 2012), so the estimates of $Q_m/Q$ should be less than the actual $Q_m/Q$. The maximum size of this bias may be assessed for the Ivory basin. When the previous estimates were made, the Ivory Glacier was undergoing significant retreat, with the contribution of flow from the glacier retreat given as 9%, almost half of the total meltwater contribution to flow (21%) (Anderton and Chinn, 1978). This may be considered an extreme example of the effect of glacial storage change. More generally, the recession of glaciers in New Zealand over the last 30 years has been at an average rate equivalent to 10 m$^3$ s$^{-1}$ (Chinn et al., 2012), which amounts to 0.1% of the South Island’s annual stream flow (Statistics NZ, 2010). Most of this comes from the Tasman Glacier, which has previously been estimated to be losing mass at an average rate of 4.3 m$^3$ s$^{-1}$ (Purdie and Fitzharris, 1999). This, however, is offset by the growth in the pro-glacial lake, so that the contribution of the glacier recession to stream flow between 1986 and 2000 is estimated at 3 m$^3$ s$^{-1}$ (Kerr, 2009). The mean annual flow of water from the Tasman Glacier catchment is estimated at 55 m$^3$ s$^{-1}$ (Woods et al., 2006) and the $Q_m/Q$ is 32%, so adding in the long-term loss of ice storage would increase this to 37%.

The partitioning of daily precipitation into snow and rain based on a threshold of daily temperature affects the results. By using a basic threshold technique, those days with a portion of the day warmer than the threshold, even though the daily temperature is cooler than the threshold, will lead to an over-estimate of snow. Similarly, those days with a portion cooler than the threshold, even though the daily temperature was warmer than the threshold, will have an under-estimation of snow. It could be expected then that regions with temperatures generally warm with respect to the threshold temperature will have an over estimation of snow. A method of assessing this error is through comparing the estimates at a site to thresholds using high-resolution climate data. Unfortunately in New Zealand there are no publicly available long-term sites that collect high-resolution snow and rain measurements. Sites do exist that have high resolution rainfall data. By using one of these sites, and altering the threshold to fit within the range of observed temperatures, the threshold effect may still be assessed. This has been done for the Greymouth Aero climate station. This site may be considered to be in the Southern Alps climate zone, and very rarely receives snow, so that the precipitation record is not affected by undercatch during snow events. Table 3 and Figure 7 show the difference between daily thresholds and 10-minute thresholds for a variety of thresholds at Greymouth Aero climate station.

If the 10-minute threshold technique is considered the truth, then thresholds based on daily temperature (average of daily minimum and maximum) leads to an underestimation of sub-threshold precipitation when the threshold is below the mean temperature for the station, and an over estimation for thresholds above the mean. The importance of the effect depends on whether the relative or absolute amounts of precipitation matter.
Table 3 – Comparison of methods to derive precipitation that occurs at the Greymouth Aero climate station below a range of temperature thresholds. See text for details of the three methods: 10 minute climate data, daily temperature and precipitation, cumulative distribution function (CDF) of a sinusoidal-linear daily temperature function.

<table>
<thead>
<tr>
<th>Threshold (°C)</th>
<th>Average annual sub-threshold total (mm)</th>
<th>Average annual sub-threshold total relative to 10 minute values (%)</th>
<th>Difference between average annual sub-threshold totals to 10 minute values (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>0</td>
<td>0 1838</td>
<td>0 2</td>
</tr>
<tr>
<td>3</td>
<td>4</td>
<td>0 155</td>
<td>–4 2</td>
</tr>
<tr>
<td>4</td>
<td>11</td>
<td>11 139</td>
<td>–10 4</td>
</tr>
<tr>
<td>5</td>
<td>27</td>
<td>20 136</td>
<td>–22 10</td>
</tr>
<tr>
<td>6</td>
<td>58</td>
<td>59 129</td>
<td>–24 17</td>
</tr>
<tr>
<td>7</td>
<td>124</td>
<td>62 122</td>
<td>–48 28</td>
</tr>
<tr>
<td>8</td>
<td>221</td>
<td>66 119</td>
<td>–76 43</td>
</tr>
<tr>
<td>9</td>
<td>361</td>
<td>88 119</td>
<td>–45 70</td>
</tr>
<tr>
<td>10</td>
<td>571</td>
<td>105 109</td>
<td>28 51</td>
</tr>
<tr>
<td>11</td>
<td>816</td>
<td>104 102</td>
<td>34 15</td>
</tr>
<tr>
<td>12</td>
<td>1038</td>
<td>103 101</td>
<td>36 10</td>
</tr>
<tr>
<td>13</td>
<td>1302</td>
<td>105 98</td>
<td>64 –24</td>
</tr>
<tr>
<td>14</td>
<td>1565</td>
<td>102 96</td>
<td>27 –63</td>
</tr>
<tr>
<td>15</td>
<td>1819</td>
<td>97 95</td>
<td>–48 –86</td>
</tr>
<tr>
<td>16</td>
<td>2041</td>
<td>98 94</td>
<td>–33 –113</td>
</tr>
<tr>
<td>17</td>
<td>2226</td>
<td>100 94</td>
<td>2 –123</td>
</tr>
<tr>
<td>18</td>
<td>2330</td>
<td>101 95</td>
<td>15 –109</td>
</tr>
<tr>
<td>19</td>
<td>2372</td>
<td>100 97</td>
<td>0 –67</td>
</tr>
<tr>
<td>20</td>
<td>2381</td>
<td>100 99</td>
<td>0 –30</td>
</tr>
<tr>
<td>21</td>
<td>2381</td>
<td>100 100</td>
<td>0 –11</td>
</tr>
<tr>
<td>22</td>
<td>2381</td>
<td>100 100</td>
<td>0 –3</td>
</tr>
</tbody>
</table>
The over estimation for thresholds of 13°C (slightly over the mean) is 64 mm, which is a lot of water, but is only a 5% error, whereas the 76 mm error for the 8°C threshold is a 34% error. Only when the threshold is near the maximum observed, do the threshold techniques converge. This result indicates that in areas where the mean temperature is significantly higher than freezing (i.e., most of the South Island), the threshold technique will underestimate sub-threshold precipitation. A potential improvement to the threshold technique is to estimate the daily variation in temperature based on the observed maximum and minimum temperatures and the known time of sunrise and sunset (Waichler and Wigmosta, 2003; Safeeq and Fares, 2011). A trial was made of estimating the variation of temperature through the day using a combination of sinusoidal and linear variation following Waichler and Wigmosta (2003). The temperature was assumed to be at the daily minimum at sunrise, varied sinusoidally to the daily maximum at 2 hours past midday, then changed linearly from sunset to the next day’s minimum on the following sunrise. An example of the result of this disaggregation for a maximum of 20°C, minimum of 12°C, sunrise at 6 am and sunset at 6 pm is shown in Figure 8.

Partitioning precipitation based on the cumulative distribution function derived from this daily temperature variation led to no general improvement (see Table 3), with overestimates for low thresholds and underestimates for high thresholds. This may be a result of temperature variation on days with rain being unrelated to the assumed daily distribution, especially when the rain is associated with frontal conditions, as is common in the New Zealand environment. It is conceivable that the temperature disaggregation system would work with
convective rain systems, which are often short-lived and associated with predictable radiative temperature effects. Exploration of disaggregation of temperature on rain days, as separate from all days and how the origins of the rain affect this disaggregation provides an interesting avenue for further investigation.

In addition to the threshold technique, the snow rain temperature threshold used for dividing precipitation into snow or rain will affect the results. In degree-day snow melt models it is not uncommon for the snow rain temperature threshold to be set above 0°C (Martinec and Rango, 1986). This is partly because snow can fall at these temperatures (Minder et al., 2011), and partly it is a tuning parameter to account for biases in the input data and process simplification (Fitzharris and Garr, 1995). Reducing the snow rain temperature threshold may also be applied as a simple approach to assessing the effect of climate warming, in that it is effectively the same as increasing all of the input temperatures. Sensitivity analysis of the snow rain temperature threshold was investigated for the Lake Pukaki and Fraser catchments (Fig. 2). $Q_m/Q$ was assessed for 7 different temperature thresholds ranging from -3 to 3°C in 1°C steps. Changing the snow rain temperature threshold resulted in a change in $Q_m/Q$ of 4.8% °C⁻¹ for Lake Pukaki, and 3.7% °C⁻¹ for the Fraser catchment (Fig. 9).

The sensitivity of $Q_m/Q$ for every sub-catchment of the Lake Pukaki area is shown in Figure 10. For low values of $Q_m/Q$ changes in the snow rain temperature threshold have a strong effect, which reduces as $Q_m/Q$ gets larger, until the temperature threshold effect becomes predominantly an offset.

This sensitivity may be explained by the relative change in catchment area that has snowfall as the snow rain temperature threshold is varied. When the $Q_m/Q$ is low, any change in the temperature threshold will have a large effect on the relative snow-covered area of the catchment. For high $Q_m/Q$ catchments, a change in the temperature threshold will have a relatively smaller impact on the overall catchment.

As mentioned above, sensitivity to the snow rain temperature threshold may be used as a proxy for sensitivity to climate change,
Figure 9 – Sensitivity of $Q_m/Q$ to the snow-rain threshold for the Lake Pukaki and Fraser catchments.

Figure 10 – Lake Pukaki river reach $Q_m/Q$ estimates, with the snow-rain threshold set to 0°C vs $Q_m/Q$ for different threshold temperatures. Each dot in each colour represents a different river reach.
in that moving the temperature threshold is the equivalent to changing the entire temperature series in the opposite direction. From Figure 10, catchments with $Q_m/Q$ of 10% given snow rain temperature threshold $= 0$ are estimated to have a reduction in $Q_m/Q$ of 3 – 5% if all daily temperatures increased by 1°C. This provides a useful general assessment of how temperature may affect snowmelt contributions to stream flow, but should be used cautiously as it does not include consideration of the specifics of how temperature may change (e.g., change in seasonality or extremes) nor does it consider how precipitation may change.

The reliability of the $Q_m/Q$ estimates depends largely on the quality of the input data. This was tested by using alternative input data. A variation on the temperature data was used whereby the interpolation to the 5-km grid from the station data was achieved using a constant lapse rate (0.005 °C m$^{-1}$), termed the ‘BA5’ temperature data (Clark et al., 2009). The lapse rate originates from observations in the Franz Josef region (Anderson et al., 2006) and matches that identified in the Aoraki/Mt Cook region (Kerr, 2005) and for the 1951–1980 mean monthly temperature normals at 301 stations throughout New Zealand (Norton 1985). It is a close approximation to the moist adiabatic lapse rate for 20°C, 1013.25 hPa. The precipitation was then varied by applying a correction surface to the daily precipitation data prior to calculating accumulations in the river channels. The correction surface, when multiplied by the mean annual precipitation enables closure of the mean annual water balance using observed flows and calculated evapotranspiration (Woods et al., 2006). The estimates for the Fraser and Pukaki catchments using the different input data combinations are shown in Table 4.

The estimates indicate that the results are sensitive to the temperature input data, but relatively insensitive to the precipitation input data. This is further demonstrated by plotting the Norton temperature, standard precipitation $Q_m/Q$ for every river reach within the Lake Pukaki catchment against the three input data cases (Fig. 11).

The BA5 temperature data decreased $Q_m/Q$ by an average of 7% when the default $Q_m/Q$ was above 10%, whereas the corrected precipitation has little overall effect on $Q_m/Q$. The effect of the BA5 temperature data is because the lapse rate used by the BA5 system for interpolating (0.005°C m$^{-1}$) is lower than average of the Norton spring lapse rates. This leads to higher freezing levels at the time of year that the extent of snowfall is most sensitive to temperature. This decreases the overall snow accumulation and hence $Q_m$. The relative effect of the change in the input temperature data is reduced as $Q_m/Q$ increases. The change in the temperature input data caused an offset in $Q_m/Q$ but the relative magnitudes of $Q_m/Q$ were largely maintained, so that the high $Q_m/Q$ catchments remained high, the low ones remained low, ensuring that the estimates remain robust as a relative measure of meltwater contribution to stream flows.

The insensitivity of the results to the precipitation input is a result of $Q_m/Q$ being

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Norton temperature, standard precipitation</th>
<th>BA5 temperature, standard precipitation</th>
<th>BA5 temperature, corrected precipitation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fraser</td>
<td>22 %</td>
<td>12 %</td>
<td>12 %</td>
</tr>
<tr>
<td>Lake Pukaki</td>
<td>23 %</td>
<td>16 %</td>
<td>17 %</td>
</tr>
</tbody>
</table>

Table 4 – $Q_m/Q$ for Lake Pukaki and Fraser catchments using the Norton temperature, BA5 temperature, standard precipitation and corrected precipitation input data.
Figure 11 – $Q_m/Q$ for each river reach in the Lake Pukaki catchment using three different input data combinations (Norton temperature with the standard precipitation, black, BA5 temperature with the standard precipitation, blue, and BA5 temperature with the corrected precipitation, red), plotted against $Q_m/Q$ using the Norton temperature and standard precipitation input data.

derived from a ratio of snow fall to total precipitation (equation (10)). Any change in total precipitation magnitude is reflected in a change of snow fall, keeping the ratio stable.

The effect of resolution of the input data on the estimates has also been tested. The high resolution precipitation field is poorly constrained in the South Island, except in that the primary control on distribution is the distance a location is from the principal orographic divide, which may be seen by the strong west-to-east precipitation gradient (Fig. 1). Previous attempts to interpolate New Zealand precipitation based on elevation have met with little success (Tait et al., 2006), precluding its use as a covariate. Creating a high-resolution precipitation field was therefore a case of resampling to a higher resolution using a bivariate spline to maintain the continuous surface. Generating a higher resolution temperature surface is possible through the close association between temperature and elevation. High-resolution daily minimum and maximum temperature data were created from the BA5 temperatures by calculating mean sea level temperature at all locations on the original 5-km grid using the 0.005°C m$^{-1}$ lapse rate and the mean elevation for each grid square (the BA5 temperatures were used instead of the Norton temperatures as it greatly simplifies the scaling of the input data). These mean sea level temperatures were interpolated using a bivariate spline to a 1-km grid. The surface elevation temperatures for these grid locations were calculated using the 0.005°C m$^{-1}$ lapse rate and elevations from a 1-km elevation grid that had been down sampled from the 30-m elevation model.

Interpolation of the input data at the daily time step for the entire South Island requires a significant increase in processing effort currently unavailable, but assessment at the catchment level provides an indication of the extent to which resolution may affect the results. To this end, the resolution tests were carried out for the Lake Pukaki and Fraser catchments.
A graph of the $Q_m/Q$ for all the river reaches in the Lake Pukaki and Fraser catchments using the 5-km input data and the 1-km input data is shown in Figure 12. The sensitivity tests show that changing the resolution of the input data from 5 km to 1 km increases the spread of the $Q_m/Q$ estimates and increases the estimates from an average of $Q_m/Q = 7.8\%$ for all of the Fraser river reaches, to 8.2\%, and from an average of 11.5\% to 12.9\% for the Lake Pukaki reaches, i.e., increases the estimates by up to a tenth.

For $Q_m/Q$ to be equal to $P_s/P$ requires that evapotranspiration is divided between the water sources in the same ratio. Determining the true partitioning of evapotranspiration is difficult, in that it requires a full understanding of soil water residence time and flow pathways and the interaction of these from melt water and rainfall. Conceptually, the release of water from a snow pack in spring ensures there is a long period of soil moisture availability for evapotranspiration processes, but potential evapotranspiration is greater in the summer months when water sources are biased towards rainfall. Spatial variability also affects evapotranspiration sources, with rainfall events covering much greater areas than snowfall (which are usually limited to higher elevations), though meltwater may be available to plants at a distance from the melt source depending on the flow paths within the soil. Determining the relative importance of these factors is non-trivial. To assess the uncertainty associated with the assumption of $ET_s/ET = P_s/P$, the range of $Q_m/Q$ for different evapotranspiration partitioning was determined. This was done by calculating $Q_m/Q$ from (8) for the Lake Pukaki and Fraser catchments, with ratios of $ET_s/ET$ ranging from 0.2–5 of $P_s/P$ and with $ET$ set at 700 mm for the Lake Pukaki catchment (McKerchar and Pearson, 1997) and 341 mm for the Fraser catchment (Fitzharris and Grimmond, 1982). The effect of varying the source of evapotranspiration from being meltwater-biased to rainwater-biased had the greatest impact in the low precipitation regions where evapotranspiration can dominate the water balance. In the Fraser catchment, $Q_m/Q$ varied from 25\% to 0\% depending on how evapotranspiration was partitioned. In the Lake Pukaki catchment the effect of evapotranspiration was less.
extreme, with $Q_m/Q$ ranging from 19% for the condition where $ET_s/ET$ was set to just 0.2 of $P_s/P$, down to 11% when $ET_s/ET$ was set to $5 \times P_s/P$.

Overall, glacier recession may be considered insignificant except in heavily glacier-dominated catchments, where it may increase the melt water contribution by a tenth. While the temperature threshold technique clearly affects the results, finding a method that improves on a simple single daily temperature threshold requires further work. The threshold temperature affects $Q_m/Q$ up to 8% °C$^{-1}$, though the larger $Q_m/Q$ amounts remain consistent from a relative perspective. Varying the input precipitation data has little effect on the results, though varying the input temperature does. The use of the Norton temperature data returned results consistent with previous estimates, while the BA5 temperature data shifted the estimates down by 7%. Reducing the resolution of the input data increased the spread of the data and increased the average $Q_m/Q$ by a tenth.

Varying the assumptions of evapotranspiration had the greatest impact in areas with limited precipitation, to the degree that all meltwater could be lost to evapotranspiration. These sensitivity assessments highlight the relative importance of the constituent parts of the estimates presented here, but do not provide any indication that the estimates should be several times larger than they are, which is what would be required to bring the estimates into line with other mid-latitude mountain regions of the world.

According to Barnett et al. (2005), the Southern Alps of New Zealand is one of the snowmelt-dominated regions of the world, alongside most of the earth above 45° latitude and mountain regions across the rest of the world. The criterion for this classification was that annual accumulated snowfall in their 0.5° latitude/longitude gridded model was greater than half the mean annual runoff. Their estimate for the Southern Alps is clearly much higher than any previous estimate for the region, but it still provides an indication of how the South Island snowmelt is thought to fit into the global picture, and is supported by other global analyses of snowmelt/runoff ratios (Adam et al., 2009). This work shows that the South Island snowmelt is significantly lower than these global analyses would suggest and considerably smaller than that reported for many similar areas of the world. For instance in the upper reaches of the Danube in Austria, with a catchment area of 77,000 km$^2$ (roughly equivalent to the combined regions of Southland and Otago), Weber et al. (2010) estimated a snow and ice melt contribution to stream flow of 27%. Similarly, the Sacramento watersheds in California (44,000 km$^2$) have been assessed as having 38% snowmelt runoff (Roos, 1991). Such melt contributions are only matched in the South Island in some of the highest alpine basins. An even more extreme example is the Maipo River (5000 km$^2$), a mountainous river basin upstream of Santiago, Chile, where the estimated average snow melt in the years 1981–1986 (766 mm) is equivalent to 75% of the mean annual precipitation (1030 mm) (Peña and Nazarala, 1987). Even the Ganges in India (917,444 km$^2$) is estimated, using satellite imagery, to have a 9% snowmelt contribution (Seidel and Martinec, 2001), three times more than assessed here for the South Island and similar to the much smaller Waimakariri catchment. All of these regions are continental with strong seasonal temperature and precipitation variability, the precipitation occurring mainly in winter, and preferentially in their respective mountain regions. These factors result in maximising the importance of snowmelt to their stream flows. By contrast, in the South Island, while snowmelt occurs, it is nowhere near as significant as in many regions of the world. This highlights how the New Zealand hydrological regime differs from that elsewhere in the world, and may be considered an outlier.
on the hydrological regime spectrum and in a unique position to help generalise findings from international research originating from continental regimes.

Conclusions
A new South Island assessment of the contribution of meltwater to mean annual stream flows has been prepared. The new assessment enables identification of the melt water portion of stream flows for any reach in the country, including at bridge, irrigation take and lake locations.

A consistent approach was applied to all river reaches, enabling simple comparisons between locations without recourse to the underlying methodology, as was previously the case. The values obtained are similar to previous estimates when the ‘Norton’ temperature input data are used. The estimates are sensitive to the temperature input data, the snow/rain temperature threshold and the partitioning of evapotranspiration between snowmelt and rainwater sources. The assessment lends itself to further development as computing capabilities enable higher resolution processing, input data are updated and/or improved (e.g., the River Environment Classification is currently being updated – Ude Shankar pers. comm. Jan 2013) and with incorporation of an appropriate temperature distribution function for precipitation thresholds.

Based on the assessment, 3% of the South Island’s mean annual stream flow is derived from snow and ice melt. This is considerably less than estimates for mountain regions of California, Austria, India and Chile. This difference is considered to be primarily a reflection of the relative lack of seasonality in precipitation and the generally mild winter temperatures. This low meltwater contribution is problematic for irrigation schemes in that artificial storage management is required if winter precipitation is to be retained for summer plant growth, however it is beneficial for hydro-electricity generation in that stream flows continue throughout the year, including winter (albeit often reduced) when energy demands are often high.

The melt percentage data is available as an Excel™ spreadsheet at: http://dc.niwa.co.nz/niwa_dc/srv/en/metadata.show?uuid=02f3e247-2f78-562a-8712-675b4f3a0166.

Acknowledgement
This work was carried out under a Ministry of Business, Innovation and Employment New Zealand Science and Technology postdoctoral fellowship. Much assistance and enthusiasm was provided by colleagues at NIWA.

References


LINZ 2000: NZMS260 series 1:50000 digital topographic map data, Land Information New Zealand (previously Department of Survey and Land Information).


Peña, H.; Nazarala, B. 1987: Snowmelt-runoff simulation model of a central Chile Andean basin with relevant orographic effects. *Large Scale Effects of Seasonal Snow Cover*, Vancouver, IAHS Publ. No. 166, IAHS.


Seidel, K.; Martinec, J. 2001: Snowmelt contributions to runoff in an extremely wide altitude range from large area satellite imagery. 5th International Workshop on Application of Remote Sensing in Hydrology, Montpellier, France.


82