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Annual ice volume changes 1976–2008 for the New Zealand Southern Alps

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ABSTRACT

New Zealand has a long, continuous record of annual end-of-summer-snowline measurements for a set of Southern Alps 'index glaciers' from 1977 to present. These index glaciers are used to estimate annual mass balance and volume water equivalent changes for the over 3000 glaciers on the Southern Alps. Two methods are employed to monitor ice volume changes. Method I deals with the rapid to normal response time glaciers, which tend to be small to medium in size. It uses mass balance gradients and glacier areas to convert changes in snowlines to changes in ice volume water equivalent. Ice volume changes for the period 1976–2008 are calculated for each index glacier, and then extrapolated to most other glaciers of the Southern Alps, using the New Zealand glacier inventory. Method II deals with 12 protracted response glaciers, which tend to be large in size. These have been slow in reacting to a long-term regional warming trend. Instead they still largely retain the ablation areas of a century ago and are in a state of disequilibrium with the present climate. These valley glaciers have recently sustained substantial ice losses that are not able to be detected using Method I. Mass balance deficits and ablation from the 12 large protracted response glaciers are estimated using a geodetic approach based on topographic and lake changes determined from repeated surveys. Results show that estimated ice volume (in water equivalents) for the Southern Alps has decreased from 54.5 km^3 in 1976 to 46.1 km^3 by 2008. This equates to a rate of $-0.3\ km^3a^{-1}$ over the last three decades, but this is considerably less than the rate of ice volume loss estimated for the previous 100 years. More than 3000 small and medium-size glaciers account for just 29% of the overall ice volume loss from the Southern Alps, while 71% of the loss occurs from the 12 large protracted response glaciers. Terminus calving contributes 0.8 km^3 and down-wasting of ice tongues in the ablation zone contributes 5.2 km^3. Some preliminary results show that ice volume changes are related to changes in circulation over the New Zealand region.

1. Introduction

Monitoring of ice volume changes across mountain ranges makes a significant contribution to the Global Climate/Terrestrial Observing System (GCOS/GTOS, 1997; Haeberli, 2004). This programme provides prime indicators for the early detection of climate-related changes associated with global warming. However, glacier monitoring data are strongly biassed towards the Northern Hemisphere, with sparse information from the Southern Hemisphere. Thus observations from the mid-latitude mountains of New Zealand, which has over 3000 glaciers, are of considerable value for the global climate monitoring network (Bishop and Forsyth, 1988; Anderson et al., 2008). Monitoring of glacier volume changes that coincided with global warming trends of the last three decades can provide information within the GTN-G (Global Terrestrial Network on Glaciers) framework (GCOS/GTOS, 1997) to the GCOS/GTOS worldwide glacier mass balance network. As reviewed by IPCC, melt of alpine glaciers has made a major contribution to observed sea level rise (Bindoff et al., 2007; Fischlin et al., 2007; and Lemke et al., 2007). Changes in glacier volume are of also of great importance for regional water resources in glaciated areas throughout the world. The impacts of climate change on water resources in New Zealand's largest river are considered by Poyck et al. (2011).

Mass balance measurements are an established method of assessing changes in glacier volume (Paterson, 1994). However, they are impractical for many countries with extensive mountain systems. Some tend to have limited financial and technical resources to devote to field glaciology and glaciers are often in remote, hard to reach locations with poor weather. Field mass balance measurements tend to be labour intensive and expensive, and even when successful provide only a small sample from just a few glaciers. Often the mass balance approach has not been able to adequately represent the myriads of glaciers that exist within an entire mountain system. For example, in New Zealand, annual mass balance measurements are published from just two glaciers, the Ivory (Anderton and Chinn, 1978) and the Tasman (Anderton, 1975; Chinn,

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Summary of units

Parameter		Unit
Annual end of summer snowline	ELA	*m.a.s.l*
ELA for equilibrium	ELA_0	*m.a.s.l*
Index glacier value	ELA_i	*m.a.s.l*
Southern Alps value	ELA_{Alps}	*m.a.s.l*
Accumulation area	S_{ac}	m^2
Ablation area	S_{ab}	m^2
Accumulation area ratio	AAR	
AAR for equilibrium	AAR_0	
Mass balance	b	*mm w.e.*
Mass balance gradient	db/dH	$mm\ m\ -^1$
Year	t	
No of index glaciers	n	
Altitude	H	*m*
Thickness, depth	h	*m*
Downwasting change	Δh	*m*
Volume loss from downwasting	ΔV	km^3 *w.e.*
Glacier ice density	γ	$890\ kg\ m^{-3}$
Mean subsectional area between surveys	S_h	m^2
Ice volume lost to lakes	V_l	*kg w.e.*
Lake area	S_l	m^2
Lake mean depth	h_l	*m*

1994), supplemented with more recent, but short-term information for the Franz Josef (Anderson, 2003; Anderson et al., 2006, 2010) and Brewster glaciers (Stumm, 2011). World-wide, mass balance measurements have been made on only about 230 glaciers (WGMS, 2008) and continuous measurements extending over periods of 20 years on only about 40 of these (Dyurgerov and Meier, 2005). Length changes are better catalogued, where there have been some 36,000 observations on 1800 glaciers.

Lack of direct mass balance measurements has meant that there are few estimates of temporal changes in glacier ice volumes for whole mountain systems. There is a clear need to find alternative ways of estimating mass balance and glacier volume changes for whole mountain systems. Rignot et al. (2003) have assessed changes for the Patagonian ice fields of South America and the contribution these have made to sea level rise. Other examples include parametric estimates of changes using multiple data sources for the European Alps (Zemp et al., 2006), New Zealand (Hoelzle et al., 2007) and Norway (Nesje et al., 2007; Baumann and Winkler, 2010). A second need is to extend these studies so as to find more cost effective, indirect ways of assessing glacier volume changes for large parts of the world. The New Zealand record of annual snowline measurements for the Southern Alps since 1977 (Chinn et al., 2005c) is used here to illustrate how these needs might be met. The aim of this study is to estimate glacier mass balances and changes in volume of ice for the whole of the New Zealand Southern Alps over the past three decades.

The bulk of the ice (53.2 km^3) in New Zealand resides along the Southern Alps of the South Island from between 42° 08′ S to 45° 56′ S and 167 ° 11′ E to 173° 38′ E. An early estimate for the total volume of water stored as perennial snow and ice for New Zealand suggested 63 km^3 (Anderton, 1973). Chinn (2001) subsequently compiled a complete glacier inventory that catalogued a total of 3144 glaciers across the country, with an estimated ice volume water equivalent (*w.e.*) of 53.3 km^3 as of 1978. Hoelzle et al. (2007) used the inventory data with theoretical parameters to estimate a volume change of −49% since the '1850 extent' to a calculated ice volume of 67 km^3 at the time of the inventory.

Wholesale retreat of New Zealand glaciers over most of the 20th century is generally attributed to regional warming (e.g. Harrington, 1952; Salinger et al., 1983; Gellatly, 1985). However, there have been reversals within this overall recession from glacier retreat to glacier advance. These have sometimes been related to regional precipitation increases (e.g. Suggate, 1950; Soons, 1971; Hessell, 1983; Brazier et al., 1992; Fitzharris et al., 1992; Hooker and Fitzharris, 1999). Hessell (1983) concluded that variations in the terminus position of the Franz Josef Glacier could be accounted for by variation primarily in precipitation amounts. Salinger et al. (1983), on the other hand, found that at the Stocking Glacier, temperature variation was the primary independent variable explaining terminus variation. Gellatly and Norton (1984) examined the apparent tension between the findings of these two pieces of work, and attributed it to differences in interpretation of meterological information. In a related process study, Kirkbride (1995) found that most measured ablation variation at the Tasman glacier could be accounted for by temperature variation. Woo and Fitzharris (1992) successfully reconstructed mass balance variation at the Franz Josef Glacier using a model based on both temperature and precipitation data and Anderson and Mackintosh (2006) found that temperature is the dominant control on glacier length. In a broader sense, response of glaciers in the Southern Alps of New Zealand is controlled by large scale changes in atmospheric circulation (Fitzharris et al., 1992, 1997; Tyson et al., 1997; Hooker and Fitzharris, 1999; Clare et al., 2002; Fitzharris et al., 2007).

2. Concepts

Two empirical methods are employed in this study to estimate mass balance and the volume of ice lost from the New Zealand Southern Alps over the past three decades. Method I is applied to the rapid to normal response time glaciers, which tend to be small to medium in size and include most of the glaciers of the Southern Alps. The method uses changes in altitude of the annual "glacier snowline" to estimate mass balance changes. Note that some large glaciers, such as the Franz Josef and Fox, are included in this methodology, because they have very fast response times. Both are largely debris free and are unusual because they descend to very low elevations of about 425 m. Method II is applied to 12 protracted response glaciers. These refer to the set of large, low gradient valley glaciers that: (a) have long response times of up to a century or so; (b) carry a thick debris mantle that insulates their tongues; (c) tend to be out of equilibrium with the present climate largely because insulation of ablation areas has delayed their retreat; and (d) are subject to a tipping point once rapid frontal retreat is initiated by lake growth. This method takes into consideration downwasting and proglacial lake development. All ice changes in this paper are expressed as water equivalent volumes.

For Method I, the "glacier snowline" as used here is measured as the altitude of the end-of-summer snowline (EOSS). The EOSS data (expressed as ELA) provides an effective proxy of the annual equilibrium line altitude if measured at the end of summer (Braithwaite, 1984; Kuhn, 1984; Paterson, 1994; Leonard and Fountain, 2003). In New Zealand, annual *ELA* values are obtained from oblique aerial photos of 50 index glaciers (Fig. 1) systematically taken from light aircraft flights over the Southern Alps in March (during the onset of austral autumn and near the end of the mass balance year). The position of *ELA* is much more easily identified on these high resolution photographs than on satellite images, particularly for the subtle multiple snow and firn lines that occur in negative balance years (Fig. 2).

These *ELA* values now provide a record of the glacier response to climate changes over nearly three decades (Chinn et al., 2005a) with very few gaps in the record. The flights are determined by the weather and are aimed for the end of the balance year. Ideally, this is the "last clear day before the first winter snowfall". Choosing the correct time to fly the stormy Southern Alps is a difficult business. Fly too soon and part of the ablation season will be missed. Fly too late and fresh new snow may obscure the EOSS and signal that the next accumulation season has begun. These decisions are based on careful monitoring of weather patterns and considered input of several experienced mountain people.

The selected 'index' glaciers are aligned on six transects that collectively span 750 km across the long axis of both sides of the Southern Alps on the South Island of New Zealand (Fig. 1). The sample set therefore includes the wetter western region that faces into the prevailing circulation direction, and also drier eastern leeward side of the Southern

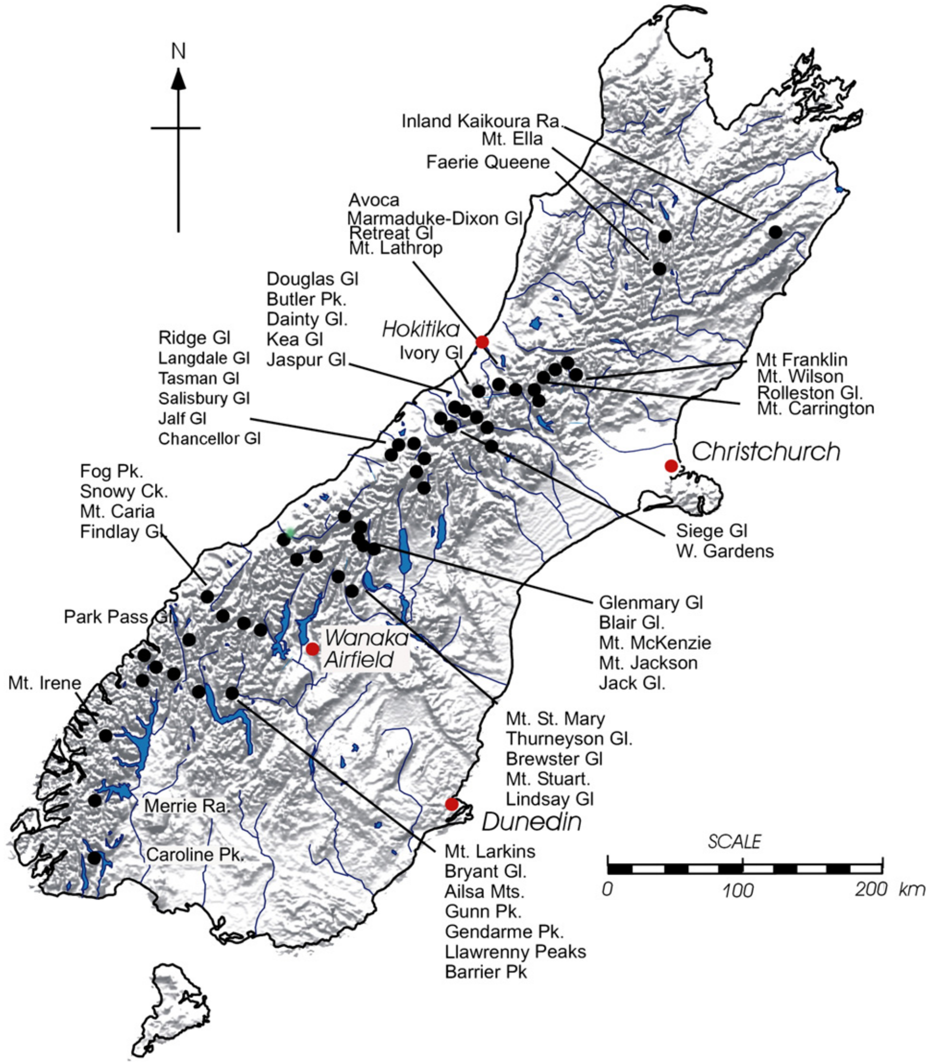

Fig. 1. Map of South Island of New Zealand showing the index glaciers. *Western zone glaciers*: Ella, Franklin, Rolleston, Retreat, Browning, Jaspur, Kea, Dainty, Butler, Seige, Vertebrae, Chancellor, Jalf, Salisbury, Jack, Jackson, Lindsay, Stewart, Brewster, Findlay, Park Pass, Caria, Llawrenny, Gendarme, Gunn, Barrier, Irene, Merrie, Caroline. *Eastern zone glaciers*: Faerie Queene, Carrington, Avoca, Marmaduke, Douglas, Butler, Tasman, Langdale, Ridge, Blair, McKenzie, Glenmary, Thurneyson, Snowy, Fog, Bryant, Larkins, Ailsa.

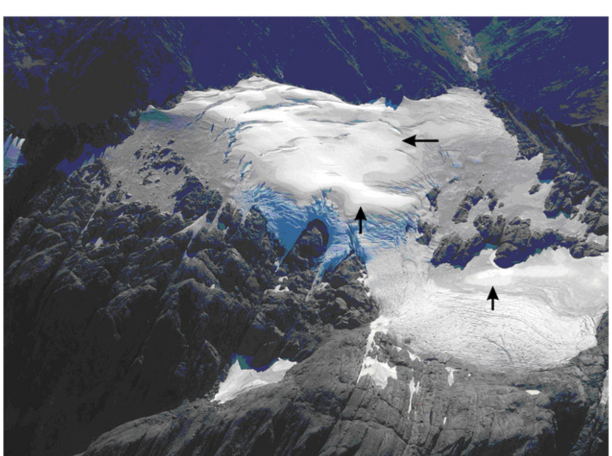

Fig. 2. Findlay Glacier, an index glacier which is in near equilibrium with the current climate. Photo taken 14 March 2008 and is typical of a negative balance year with a high end-of-summer snowline (arrows) occurring as isolated remnant snow patches above numerous old firn lines.

Alps; (Chinn, 1979; Griffith and McSaveney, 1983). Wherever possible, the index glaciers were chosen for their simplicity of form, with even slopes about the equilibrium line altitude (see Chinn, 1995 for details). The fluctuations of *ELA* are published annually (e.g. Chinn and Salinger, 1999; Chinn et al., 2005a). While there are small glaciers on the North Island, they are not considered here because there have been no systematic EOSS surveys of these glaciers.

Rabatel et al. (2005) have demonstrated how *ELA* can be used for the calculation of mass balance, as described for three French glaciers. Their methodology is extended to the annual *ELA* measurements for index glaciers of the Southern Alps of New Zealand to derive estimates of mass balance changes. When linked to the New Zealand glacier inventory, changes in ice volume *w.e.* can be estimated for the whole of the Southern Alps, provided large valley glaciers are treated separately as in Method II below. Preliminary results from an elementary version of Method I were first obtained by Chinn et al. (2002).

Timing of the EOSS survey programme has been fortuitous because, contrary to world-wide mass balance trends, the New Zealand data have displayed a cumulative mass balance that is near zero for the last three decades (WGMS, 2008). Nevertheless, inter-annual oscillations of both positive and negative mass balance over this period have provided a wide range of response conditions when compared with the more general world-wide case of constantly receding glaciers. Of further significance, most Southern Alps glaciers have had sufficient time to attain a new equilibrium area consistent with current climate. This is because the majority of the glaciers, which are of small or medium size, have many of the short response time characteristics outlined by Paterson (1994), including short steep profiles, large mass turnovers typical of high precipitation locations, and high terminus ablation rates. These factors all lead to quick adjustments to climate changes. Thus a large proportion of New Zealand glaciers have reached equilibrium with the present climate. Observed response times are between 6 and 20 years. Hoelzle et al. (2007) suggest a mean of about 12 years.

Over the last four decades, most glaciers of the Southern Alps have exhibited advances, followed by retreats and re-advances (Chinn, 1996, 1999; Chinn et al., 2005b). However, the12 protracted response glaciers maintained their “Little Ice Age” areal extent right up until the 1970s (Fig. 3). This is despite sustained warming across New Zealand (Salinger and Mullan, 1999; Mullan et al., 2010) since the late 19th century. As their ELAs rose by warming, they lost ice volume to downwasting, but largely retained their ablation zone area. This led to exceedingly small AAR_0 values, a measure of their extreme disequilibrium (Dyurgerov et al., 2009). Retention of ablation areas of these glaciers was a result of: (a) a heavy insulating layer of debris, thickening to as much as 2 m at the termini, which retarded ablation rates; (b) low surface gradients; and (c) very thick ice extending well below the melt water river outlet level, which prevented terminus retreat (see the diagram Fig. 4).

The tongues of these 12 glaciers were effectively large, active “ice ponds” lying in over-deepened hollows that could not easily retreat or even reduce size in response to the warmer climate (Fig. 5). They lost ice volume by simple down-wasting until the terminus ice levels subsided to the outlet river level, which occurred separately between the 1970s and 1990s. This critical development (Fig. 5) was a tipping point for these large glaciers. Thereafter, they experienced rapid lake expansion, catastrophic ice loss from calving (see Fig. 6) and destruction of their lower trunks (Hochstein et al., 1998; Kirkbride and Warren, 1999; Purdie and Fitzharris, 1999; Röhl, 2006; Dykes et al., 2010; Winkler et al., 2010). These ice losses were additional to those from mass balance changes associated with fluctuations in annual ELAs. Consequently, ice volume losses from large glaciers have to be assessed using an alternative Method II. It is based on geodetic and topographic changes as determined from repeated surveys.

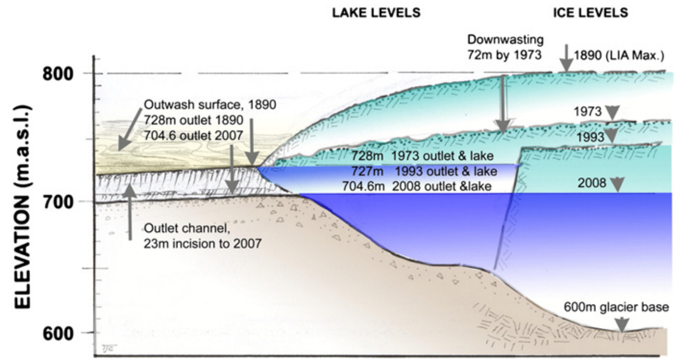

Fig. 4. Progressive stages of ice loss from a large valley glacier of the Southern Alps associated with regional warming. At the LIA, the meltwater river feeds directly on to an alluvial outwash gravel fan head. Subsequent stages are indicated: downwasting from LIA levels to formation of thermokast features on the terminus; formation of pro-glacial lake; calving and retreat of ice cliff. Dates and elevation measured at the Tasman Glacier (after Röhl, 2006; Hochstein et al., 1998).

3. Glaciological parameters

3.1. ELA and mass balance

Usually located somewhere above mid-glacier, the ELA at the end of summer indicates the altitude where snow accumulation is at a zero net gain over snow loss for the past mass balance year Meier and Post (1962) (Fig. 7). The end-of-summer snowline (EOSS), the feature measured in this project, has a value very close to that of the ELA and for the purposes of this paper is assumed to be equivalent to the ELA. The EOSS is normally visible as a line of demarcation on the glacier surface. It is visible due to the contrast between discoloured firn and ice below the EOSS (tinted brown due to dust and sediment accumulation that is a residue of ablation) and clean snow of the previous winter above the EOSS (Fig. 2). Following Meier and Post (1962), the altitude of this ‘glacier snowline’ is considered a good approximation of the annual ELA. EOSS altitudes have been documented each year for three decades by aerial snowline surveys in New Zealand (Willsman et al., 2010), and used as a proxy for annual mass balance changes at each glacier (Chinn,

Fig. 3. Hooker Glacier, April 2007. This is typical of the large valley glaciers of the Southern Alps, where recent lake growth is now rapidly consuming the terminus area. The extent of down-wasting is shown by the height of the grey moraine wall crest which marks the LIA areal extent and ice level of the glacier. The thick mantle of debris cover on the glacier tongue has persisted since LIA time and has significantly retarded down-wasting rates. View looking north towards Aoraki/Mt Cook (elevation 3754 m).

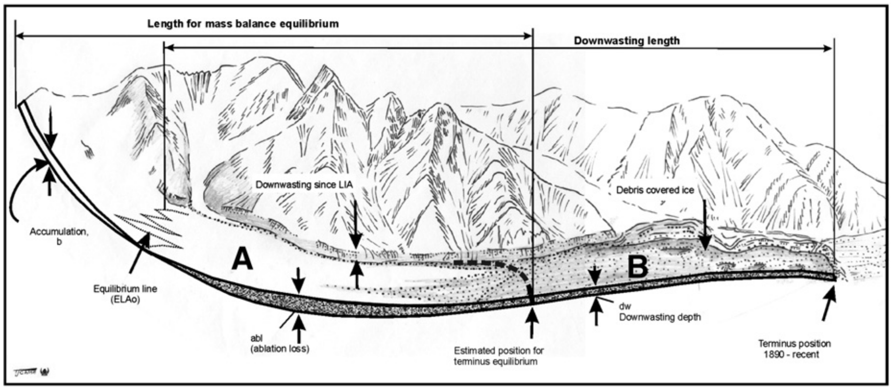

Fig. 5. Zones encountered on the large debris mantled glaciers typical of the Southern Alps. A = estimated area and terminus position of the glacier for it to be in equilibrium with the present climate. B = the actual area before any lake development and extends down to the LIA (1890) terminus position. Area 'B' may be considered a measure of the disequilibrium of the glacier and is largely debris covered 'relict' but active ice. The loss to down-wasting is calculated from the ELA_0 down to the existing terminus. Treatment of apparent double sampling of ice loss is explained in the text.

1995; Chinn et al., 2005c). 30 years of data with both positive and negative balance measurements from 50 glaciers has meant the long-term equilibrium line altitude (ELA_0) has been derived using a regression plot for each glacier (Chinn et al., 2005c), which is similar to the method used by Dyurgerov et al. (2009). Annual *ELA* for each glacier is individually regressed against the average of all other annual index glacier values measured throughout the Southern Alps (ELA_{Alps}) for each respective year. This method provides ELA_0 values for each index glacier that are unaffected by problems associated with averaging a limited number of *ELA* values. Annual differences between ELA_0 and *ELA* provide a relative measure of mass balance changes that can be converted to the departure of the annual mass balance (b) from the equilibrium mass balance (Fig. 7). This leads to quantitative estimates of mass balance change by using appropriate glacier mass balance gradients and hypsometry by following the method of Rabatel et al. (2005).

3.2. Accumulation area ratio (AAR)

Glacier studies worldwide have demonstrated that for glaciers in equilibrium, the ratio of the accumulation area to the total glacier area (AAR_0) is on average close to 0.66. This indicates a 2:1 ratio of the accumulation zone area to ablation zone area. This value has been used extensively in glaciology (e.g. for derivation of paleosnowline altitudes (Gross et al., 1978; Kasser, 1973; Maisch et al., 1999; Porter, 1975)). However, it is known that the *AAR* varies considerably with glacier morphology and AAR_0 can range between about 0.25 to 0.75 (e.g. Glacier Mass Balance Bulletin No 8, 2002–2003), with largest deviations for debris-covered and abnormally shaped glaciers, while Bahr et al. (2009) report a mean AAR_0 value of 0.57 for a sample of 87 glaciers. The ELA data gathered on the EOSS flights over the Southern Alps during the last 30 years has supplied a large dataset that suggests an AAR_0 of 0.66 or 2:1 is an appropriate average for New Zealand.

3.3. Glacier mass balance gradient

The mass balance gradient (db/dH) is the change in annual mass balance (b) as a function of the glacier altitude (H). Since the accumulation area is assumed to be twice the ablation area, the ablation area gradient is assumed to be twice that of the accumulation area (Fig. 7). While most studies use identical mass balance gradients for both accumulation and ablation areas measurements from many glaciers show that this is hardly ever the case (e.g. Glacier Mass Balance Bulletin No 8, 2002–2003). For this study, it was considered that accuracy of volume estimates can be improved by incorporating the 2:1 ratio into the mass balance gradients.

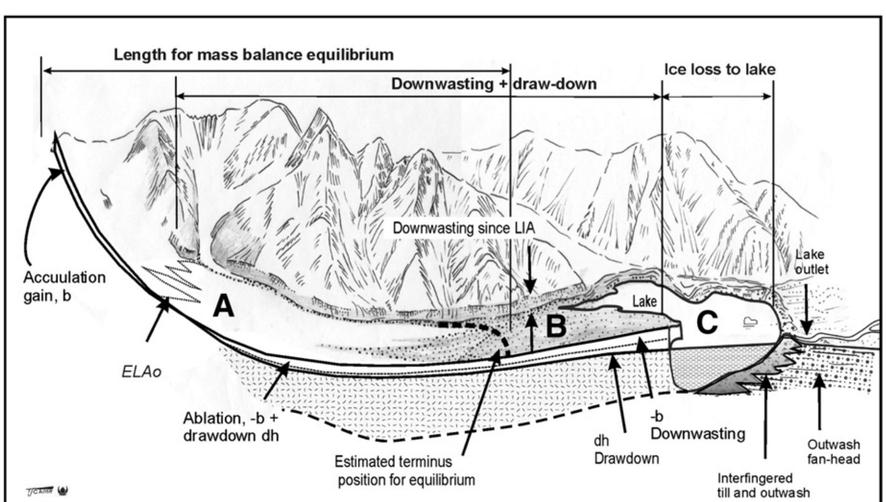

Fig. 6. Zones encountered on the large debris mantled glaciers typical of the Southern Alps after pro-glacial lakes develop. A = area which would be used for annual ice volume changes using changes in annual ELA_0. However, there are also ice volume losses from downwasting of the relict debris covered ice tongue 'B' and from calving into the expanding lake 'C'. In addition, there is a smaller loss to drawdown as the flow of ice of trunk 'B' accelerates into the lake. For these glaciers Method II is applied: repeat topographic measurements provide the values of ice volume loss from down-wasting and drawdown; measurements of increases in lake volume are used to estimate ice losses to the lake.

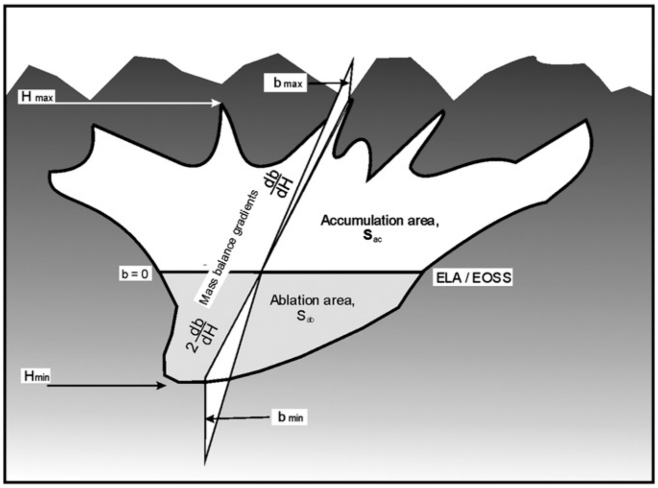

Fig. 7. Scheme explaining parameters used for calculation of annual ice volume changes for index glaciers as used in Method I.

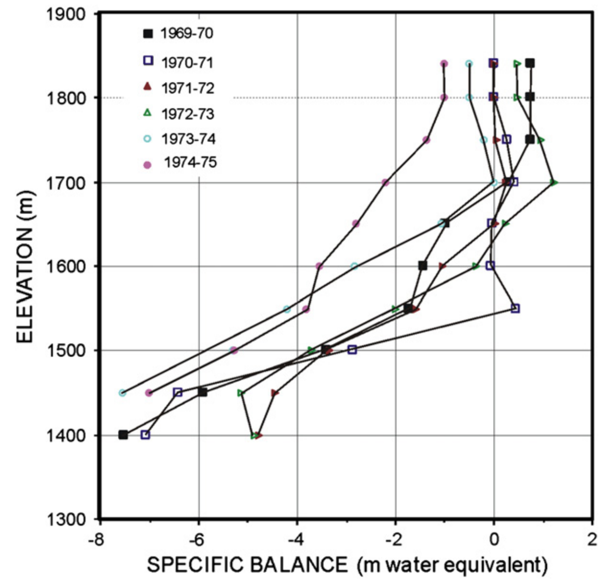

Fig. 8. Variation of mass balance with elevation for Ivory Glacier over 6 glacier years. Note the sharp decrease of the mass balance gradient (*db*/*dH*) from the ablation zone to the accumulation zone.

Mass balance gradients increase with increasing precipitation, and are more variable in wetter climates than in drier climates (Kuhn, 1981; Oerlemans and Fortuin, 1992). Mass balance gradients for continental areas are normally low, being <5 mm m^{-1} (w.e.), and at the extreme end of the range in the Antarctic Dry Valleys region, gradients reduce to 0.14 to 0.55 mm m^{-1} (Chinn, 1980). For maritime glaciers they are much higher and typically exceed 15 mm m^{-1} (Holmlund and Fuenzalida, 1995).

The largely maritime climate of New Zealand produces very high *db*/*dH*, but these values need to be adjusted for the strong west–east precipitation gradient across the Southern Alps. For example, precipitation reaches a maximum of 10–15 ma^{-1} for glaciers to the west of the Main Divide, which is upwind of the prevailing westerly circulation and declines exponentially to about 1.5 ma^{-1} over the easternmost glaciers (Chinn, 1979). There are few data for estimating mass balance gradients. Higher values are assumed for the “wet” western glaciers and lower values for the “dry” eastern glaciers, a similar approach to that used by Hoelzle et al. (2007).

Table 1 and Fig. 8 show *db*/*dH* values measured for 6 years for Ivory Glacier (Anderton and Chinn, 1978), which is located in the maximum precipitation zone. It is considered to be representative of the glaciers west of the Main Divide, although frontal calving commenced as the study progressed. However, during the six-year measurement period snowlines were almost at the top of the accumulation zone. The few accumulation zone values of *db*/*dH* available for Ivory Glacier are judged unreliable, but are estimated to be 12.9 mm m^{-1}, or half the measured ablation zone *db*/*dH* of 25.8 mm m^{-1} based on the data of Anderton and Chinn, 1978. The few measured “dry” eastern mass balance gradients are listed in Table 1 and shown in Fig. 9. Values measured for the Tasman Glacier (Anderton, 1975), are chosen as representative of the eastern region, with a *db*/*dH* of 7.5 mm m^{-1} for the accumulation zone. For the ablation zone, a value of 15.0 mm m^{-1} was used, consistent with an AAR of 0.66.

4. Methods

In their world-wide assessment of mass balance measurements for the period 1961 to 2004, Dyurgerov and Meier (1997, 2005) used water equivalent depth changes averaged over an entire glacier. Annual volume changes were calculated as the product of specific mass balance (in water equivalent) and glacier area and extended to all glaciated areas of the world, excluding the large ice sheets. A similar approach is applied here to the Southern Alps of New Zealand, where annual mass balance changes are derived from *ELA* values from index glaciers and extrapolated throughout the study area using the New Zealand glacier inventory (Method I). This inventory was compiled in 1978, (Chinn, 2001) and is used in this study as the baseline for the glacier dimensions. The set of index glaciers (Fig. 1) was chosen to form a representative sample of the Southern Alps. Their *ELAs* for each year are considered to be representative of annual ELAs for the whole glacier inventory. In

Table 1
Altitudinal mass balance gradients (*db*/*dH*) (mm m^{-1}) derived from measurements on Ivory, Tasman and Brewster Glaciers. Asterisked values are estimated (see text).

	Ivory Glacier		Tasman Glacier		Brewster Glacier	
Year	Ablation zone	Accumulation zone	Ablation zone	Accumulation zone	Ablation zone	Accumulation zone
1966–67				6.0		
1967–68				8.6		
1968–69		No	Unreliable	7.5		
1969–70	26.3			Unreliable		
1970–71	35.5	Useful	11.7	Unreliable		
1971–72	20.8		13.7	7.4		
1972–73	28.9	Data	12.4	9.4		
1973–74	28.0		14.3	7.0		
1974–75	15.4		10.4	6.6		
2004–05					19.9	6.8
2005–06					18.9	6.0
2006–07					20.8	7.6
Mean	25.8	12.9*	15.0*	7.5	19.9	6.8

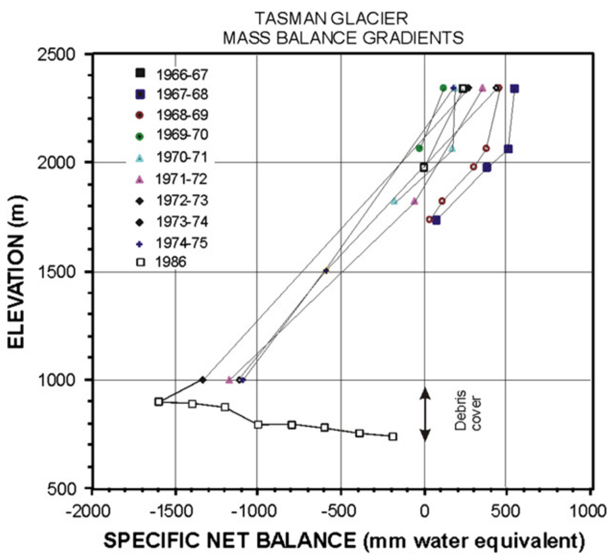

Fig. 9. Variation of mass balance with elevation for Tasman Glacier over 9 glacier years. Note gradient reversal below 900 m caused by the debris mantle which increases in thickness to >2 m towards the terminus. Data is from Anderton (1975) with 1986 measurements after Kirkbride and Warren (1999).

general the outline areas of the index glaciers have changed little since 1978, despite undergoing large inter-annual ELA changes. This suggests that they are close to equilibrium with the present climate and this simplifies volume change calculations. *ELA* measurements from the 50 index glaciers are used to calculate mass balance (b) following the methods of Rabatel et al. (2005), but modified by using separate mass balance gradients for ablation and accumulation areas (for the reasons outlined in Section 3.1). The specific mass balance is then calculated for the whole of the Southern Alps for each year (b_{Alpst}) using the annual mean specific mass balance of the index glaciers: Quantitative mass balances are then applied to all glaciers of the inventory to obtain ice volume changes for the entire Southern Alps for the period 1976–2008.

For the 12 large protracted response glaciers (Figs. 3, 4, 5, and 6) that are still adjusting to 20th century warming, Method II was applied. These glaciers retain a surface area much greater than that conformable with an equilibrium state under present climate conditions. They need to be treated separately so as to account for on-going ice losses at their termini (see reasons in Section 2). Method II is based on repeat topographic measurements and estimates of ice losses from calving into newly formed pro-glacial lakes. Integration of Method II with Method I is difficult, but is largely accommodated by assuming that the latter calculation produces a near zero cumulative mass balance change between the beginning and end of the study period. Both methods are described in detail below.

4.1. Method I—snowline and estimates of mass balance for index glaciers

4.1.1. Application to index glaciers

Volumetric water equivalent changes are calculated for each index glacier, which are supported by a considerable amount of data from small to moderate sized glaciers across the Southern Alps. The annual estimate of the *ELA* as depicted from the EOSS photographic data (i) is plotted for each year on to a contour map visually or by photo rectification, to enable measurement of both the accumulation and the ablation areas. A single ELA value for every index glacier (ELA_i) is then derived from the area-altitude curve which also expresses the glacier hypsography. This process also provides the value of the accumulation area ratio (A_i) for each index glacier.

ELA_0 was pre-obtained using regression plots from the long Southern Alps *ELA* record (see Section 3.1 and details in Chinn et al., 2005c), where (i) refers to an individual index glacier for a given balance year (t). By contrast, the annual value for the whole Southern Alps (ELA_{Alps}) can be calculated simply as the average of index glacier values for any given year (Eq. (1)):

$$ELA_{Alps} = \frac{1}{n}\sum_{i=1}^{n} ELA_{it}. \tag{1}$$

From Fig. 7, for any given year, the net snow accumulation for each index glacier is the product of mean depth (b_{ac}) and the accumulation area (S_{ac}), which are converted to cubic metres of water equivalent. Likewise, the ablation loss is the product of mean depth (b_{ab}) and the ablation area (S_{ab}). The resulting volumetric change in water equivalent is seen in Eq. (2):

$$\Delta V_{ab} = accumulation\ gain(S_{ac} \cdot b_{ac}) – ablation\ loss\ (S_{ab} \cdot b_{ab}). \tag{2}$$

For each annual value of the change in the ELA ($ELA_0 - ELA$), the respective annual areas S_{ac} and S_{ab} are read off the area-altitude curve of the particular glacier.

Mean mass balances for both the accumulation and ablation zones of each index glacier are obtained from the db/dH curves (Figs. 8 and 9). The altitude range for the glacier is from its highest point (H_{max}) to its lowest elevation (H_{min}).

For mean annual accumulation specific to an individual index glacier (Eq. (3)):

$$b_{ac} = ½\, b_{max} = (H_{max} - ELA_t) \cdot db/dH_{ac}/2. \tag{3}$$

For mean annual ablation specific to an individual index glacier (Eq. (4)):

$$b_{ab} = ½\, b_{min} = (H_{min} - ELA_t) \cdot db/dH_{ab}/2. \tag{4}$$

Use of an average AAR of 0.66 then leads to the following proportional relationship (Eq. (5)):

$$db/dH_{ab} = 2db/dH_{ac}. \tag{5}$$

The mass balance gradients are then applied to accumulation and ablation areas for the index glaciers. 12.9 mm m^{-1} and 25.8 mm m^{-1} respectively to 33 suitable glaciers in the wetter, western region (all glaciers to the west of the Main Divide) of the Southern Alps while for 13 index glaciers in the drier eastern region, (all glaciers to the east of the Main Divide) values of 7.5 mm m^{-1} and 15.0 mm m^{-1} are used (Fig. 1). The remaining 4 of the index glaciers were deemed unsuitable for this study.

For each index glacier for each balance year, the change in volumetric water equivalent (ΔV_{it}) is shown in Eq. (6):

$$\Delta V_{it} = \frac{S_{ac}(H_{max} - ELA_{it})\frac{db}{dH_{ac}}}{2} - \frac{S_{ab}(ELA_{it} - H_{min})\frac{db}{dH_{ab}}}{2} \tag{6}$$

4.1.2. Application to most other glaciers of the Southern Alps

Annual glacier mass balances are extrapolated from the index glaciers to all other small and medium sized glaciers across the Southern Alps. For each index glacier (i) the specific mass balance is calculated for each year (t) by dividing each volume change (ΔV_{it}) by the index glacier area (S_{it}) in Eq. (7):

$$b_{it} = \Delta V_{it}/S_{it}. \tag{7}$$

The specific mass balance is then calculated for the whole of the Southern Alps for each year (b_{Alpst}) using Eq. (8):

$$b_{Alpst} = \frac{1}{n}\sum_{i=1}^{n} b_{it}. \tag{8}$$

The water equivalent volume change for the Southern Alps for each balance year can then be derived using Eq. (9):

$$\Delta V_{Alpst} = S_{Alps} b_{Alpst} \tag{9}$$

where S_{Alps} is the total glacier area for small to medium sized glaciers of the Southern Alps, which annual observations show, have changed little since 1978. The aggregate of changes in ice area for all glaciers without terminal lakes is close to zero (Chinn, 1999). ΔV_{it} is then summed cumulatively for all years for all glaciers. A total ice volume *w.e.* of 53.2 km^3 and an area of 1153 km^2, as obtained for the glacier inventory in 1978, are taken as initial values for these calculations.

4.2. Method II—long valley glaciers with calving termini

The adjustment to long-term warming from the Little Ice Age is still on-going for the 12 large protracted response glaciers of the Southern Alps (see Table 2 for details). These glaciers have exceedingly small AAR_0 values and are shrinking rapidly (Dyurgerov et al., 2009). Ice loss calculations for these glaciers consider two processes. The first is "downwasting" (see Fig. 5) defined here as the net lowering of the surface profile by normal ablation and ice flow. The second incorporates the development since the 1970s of pro-glacial lakes (Figs. 4 and 6) and the combined ice volumes lost to both downwasting and calving. The effects of these expanding lakes plus the associated draw-down of the glacier surface and the consequent increases in flow velocities are all accounted for in the topographic measurements. The relevant mechanisms initiating the pro-glacial lake growth are described above in Section 2, Concepts.

The rationale for Method II is illustrated by the block diagram and cross section of a valley glacier seen in Fig. 5. In this example, the normal annual mass balance change is shown in area 'A' from the accumulation area through the equilibrium line to a hypothetical (and unknown) terminus position which would represent equilibrium to current climate. For this section of the diagram only, Method I applies, that is "Annual mass change = accumulation − ablation". However, there are considerable, additional ice losses that need to be considered for the relict area 'B' that require additional measurements. These involve repeat surveys of the glacier surface altitude between the present equilibrium line and the terminus. This approach neglects volume changes of the accumulation area, but this is considered acceptable for the larger valley glaciers because:

(1) net balance over the study period is approximately zero, so annual surface mass balance changes of area 'A' contribute little to total glacier volume changes;

Table 2
Estimates of ice volume losses (w.e.) by downwasting for the 12 large protracted response glaciers (assumes an ice density of 890 kg m^{-3}).

Glacier	Period	Mass loss due to down-wasting (w.e. km^3)	Down-wasting Δh along CFL (m a^{-1})	Method (see footnotes)	Estimated mass loss for both periods 1977–1985 and 1986–2007 (w.e. km^3)
Tasman	1964–1986	−1.30	−1.9 m a^{-1} over 10 km	a	
	1986–2007	−1.10	−2.1 m a^{-1} over 10 km	b	−1.74
Godley	1964–1986	−0.88	−3.5 m a^{-1} over 4 km	a	
	1986–2007	−0.57	−3.9 m a^{-1} over 4 km	b	−1.00
Murchison	1964–1986	−0.53	−1.7 m a^{-1} over 10 km	a	
	1986–2007	−0.70	−2.5 m a^{-1} over 10 km	b	−0.99
Classen	1964–1986	−0.17	−1.0 m a^{-1} over 3 km	a	
	1986–2007	−0.25	−1.9 m a^{-1} over 3 km	b	−0.35
Mueller	1964–1986	−0.20	−1.3 m a^{-1} over 2 km	c	
	1986–2007	−0.20	−1.4 m a^{-1} over 2 km	d	−0.27
Hooker	1964–1986	−0.10	−0.9 m a^{-1} over 2 km	c	
	1986–2007	−0.12	−1.5 m a^{-1} over 2 km	d	−0.18
Ramsay	1964–1986	−0.13	−1.5 m a^{-1} over 2 km	e	
	1986–2007	−0.11	−1.2 m a^{-1} over 1 km	f	−0.17
Volta/Therma	1964–1986	−0.09	−0.7 m a^{-1} over 4 km	e	
	1986–2007	−0.09	−0.7 m a^{-1} over 4 km	g	−0.15
La Perouse	1964–1986	−0.12	−1.2 m a^{-1} over 3 km	e	
	1986–2007	−0.06	−0.7 m a^{-1} over 1 km	f	−0.12
Balfour	1964–1986	−0.07	−1.5 m a^{-1} over 2 km	e	
	1986–2007	−0.07	−1.5 m a^{-1} over 2 km	g	−0.11
Grey	1964–1986	−0.06	−1.8 m a^{-1} over 2 km	e	
	1986–2007	−0.07	−1.0 m a^{-1} over 1 km	d	−0.11
Maud	1964–1986	−0.04	−1.2 m a^{-1} over 2 km	e	
	1986–2007	−0.02	−0.7 m a^{-1} over 1 km	d	−0.04
Total (km^3)					−5.23

Under Methods column, key is as follows:

a = Central flow line (CFL) altitude change between aerial photograph map contours (APMC) in 1964 and 1986. Surface lowering (Δh) applied to ~1 km subsections from glacier terminus to altitude of ELA_0, thence from ELA_0 to top (H_{max}) treated as zero change.
b = CFL change from 1986 APMC and 2007 GPS elevation survey to ELA_0, applied as in Method a.
c = Method a with ~2 km subsections of tongue to ELA_0.
d = CFL change from 1986 APMC to 2007 GPS survey on lower 2 km of the glacier, applied to 1 km subsections of the glacier to the altitude of ELA_0, elevation change estimated above the 2007 GPS survey by proportionally lowering the 1986 CFL profile.
e = Lower glacier (2–3 km to terminus) CFL change from 1964 to 1986 from APMC single average from zero at ELA_0 to terminus, applied to glacier area.
f = 1986 altitude from Method e and estimated 2007 altitude from 1986 lake height and aerial photograph estimate of the cliff height, single average from zero at ELA_0 to terminus, applied to the area of the glacier.
g = 1986 to 2007 assumed the previous periods rate, as terminus has retreated out of a pro-glacial lake, oblique aerial photographs show down-wasting continuing.

All 2007 values were measured in April from a GPS survey using a Trimble 5700 RTK receiver and base using NZGD2000 datum, NZTM projection, corrected to Mt John, NZ reference station (Geodetic code MTJO), and heights transformed to MSL using the NZVD05 geoid model. A GPS elevation check with geodetic mark A9A1 at Mt. Cook airport was <0.2 m from the orthometric (MSL) height. Map elevation sources refer to 1964 NZMS1 series, 1986 Topomap series, as published by Land Information New Zealand (LINZ).

(2) on-going down-wasting and calving losses are much greater than the above annual mass balance changes;
(3) insulation from the debris mantle thickening towards the terminus means that the ablation–altitude gradients are reversed (Fig. 9) and normal mass balance gradients typical of other glaciers in the Southern Alps do not apply here (see discussion of this point in Kirkbride and Warren, 1999).

4.2.1. Ice losses from down-wasting

Although the mean response time for most glaciers of the Southern Alps is short (~12 years, Hoelzle et al., 2007), the largest valley glacier responses are much greater than 80 years. These glaciers are still adjusting their size to the warming from the end of the 19th century, since New Zealand temperatures have increased by about 1 °C (Folland et al., 2003), and are still receding from late Little Ice Age terminal positions, as evidenced by well-dated moraines (Schaefer et al., 2009). The original convex parabolic longitudinal profiles for tongues of the 12 large protracted response glaciers have flattened to straight or even concave surface slopes that are commonly pockmarked by thermokast features and supra-glacial ponds. The ice volume loss from down-wasting is calculated by dividing the glacier tongue surfaces into sub-sections and delineating the glacier central flow lines (CFL). Changes of CFL elevations are measured from 1964 and 1986 reference maps and GPS elevations taken from surface surveys conducted in 2007. The 1964 to 1986 interval was arithmetically proportioned to 1977. (Note that although the first EOSS survey was made in 1977, it measured the EOSS for the 1976–77 glacial year). The fact that many of the glacier tongues are heavily mantled by debris is accommodated by these surveys. The CFL altitude changes are averaged across and applied to each glacier sub-section, with the assumption that the ice loss is consistent across the width of the glacier in each subsection. The mass (ΔV) loss is calculated by CFL altitude change (Δh), multiplied by the mean sub-sectional area between surveys (S_h), as shown in Eq. (10):

$$\Delta V_h = \gamma \Delta h\, S_h. \tag{10}$$

To convert to volume w.e., the glacier ice is assumed to have a density (γ) of 890 kg m^{-3}. The details are documented in Table 2.

4.2.2. Ice losses due to calving

Ice margin recession at the glacier terminus is calculated directly from photographic records of lake growth that have been documented since 1980 (Chinn, 2002) and from bathymetric measurements (Warren and Kirkbride, 1998; Röhl, 2006,; Hochstein et al., 1998). Detailed rates of ice calving into lakes are also available for the Hooker and Tasman Glaciers (Kirkbride, 1993; Hochstein et al., 1995, 1998; Kirkbride and Warren, 1999; Purdie and Fitzharris, 1999; Kääb, 2002; Röhl, 2006). The mechanisms of pro-glacial lake growth are outside the scope of this paper; only the observed consequences of these processes are discussed. The volume of ice lost to the growth of the pro-glacial lakes (V_l) for each balance year is derived from the product of the lake area (A_l) and lake mean depth for each year (d), shown in Eq. (11):

$$\Delta V_l = \gamma\, S_l d. \tag{11}$$

5. Results

5.1. Volume changes from snowline/mass balance fluctuations

Changes in specific mass balance for the Southern Alps for the period 1977–2008 are shown in Fig. 10 Each of the decades of the 1980s, 1900s and 2000s had clusters of consecutive positive mass balance years with values ranging from 0.5 to 2.1 m. These were years when ELA_{Alps} were below normal. These clusters are separated by occasional strongly negative mass balance years with values ranging from −2 to −4 m. w.e. These are more pronounced than the positive years and indicate ELA_{Alps} were above normal. Ice volume changes calculated using Method I (Fig. 11) show considerable inter-annual variability for the small glaciers of the Southern Alps. There were annual losses of 1 to 2 km^3 w.e. a^{-1} in glacier years 1977/78, 1978/79, 1997/98, 2005/06, and of 2 to 4 km^3 w.e. a^{-1} in 1989/90, 1998/1999, 1999/2000, 2001/02, and 2007/08. There were annual gains of 0.5 to 2 km^3 a^{-1} from 1982/83–1984/85, 1991/92–1994/95 and 2002/03–2004/05. The cumulative mass balance shows a total loss of 2.45 km^3 w.e. for the Southern Alps from small and medium size glaciers over the 1976–2008 period.

Mass losses in water equivalents for the 12 large protracted response glaciers of the Southern Alps using Method II are estimated as 0.75 km^3 for terminus calving and 5.22 km^3 for tongue downwasting over the 1976–2008 period (Table 2). This represents a total loss of almost 5.96 km^3 w.e over 32 years. This is much larger than the loss of ice from small and medium glaciers of 2.45 km^3. These calculations indicate that total ice volume of the Southern Alps has decreased from 54.53 km^3 in 1976 to 46.12 km^3 in 2008 (a loss of 8.41 km^3 or 15%) at a rate of 0.26 km^3 a^{-1} (Figs. 11 and 12).

More than two thirds (5.96 km^3) has come from just 12 of New Zealand's largest glaciers and is due to tongue down-wasting and terminus calving into expanding pro-glacial lakes. The relative masses of ice lost to the three main sources of mass balance, lake growth/ice calving and downwasting are given in the comparative losses of Fig. 13.

6. Discussion

Globally, the last three or four decades have been dominated by widespread glacier recession and substantial loss of alpine ice volume (Beniston, 2003). However, the 32-year period of index glacier surveys in New Zealand has occurred during a period of both positive and negative mass balances (Fig. 10) with associated advance and recession of many glacier frontal positions (Chinn, 1999; Chinn et al., 2005b). These fluctuations have provided two research benefits. First, they have been driven by a wide range of annual mass balances (Chinn, 1996; Hooker and Fitzharris, 1999; Chinn et al., 2005b). Second, there has been little change in areal extent of all but a few of the largest glaciers as averaged over the survey period; the vast majority of glaciers are essentially in equilibrium with the present climate. Without these factors, it would have been more difficult to relate changes in snowline altitudes to ice volume changes, compared with say, a dataset of continuous negative mass balance and glacier retreat.

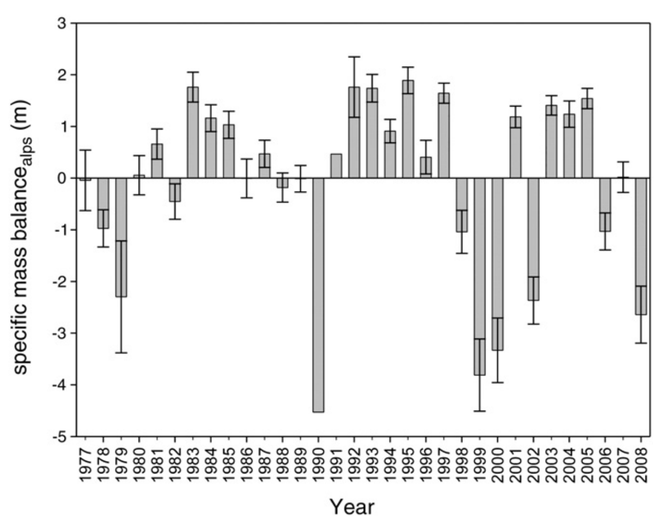

Fig. 10. Changes in specific mass balance for the Southern Alps from 1977 to 2008 for the small to medium in size rapid to normal response time glaciers. Also shown are 95% confidence limits based on specific mass balance of 50 individual index glaciers. Estimates for the 1989/90 year are from observations for only two index glaciers, and for one index glacier from 1990/91.

Fig. 11. Ice volume changes (w.e.) for the Southern Alps from 1976 to 2008. The dashed line shows results using Method I. The solid line adjusts results after applying Method II.

The ELA measurement programme for New Zealand enables sampling of ELA proxies of a substantial mountain system at the cost of less than 2 days per year of aircraft flight time (Willsman et al., 2010). Alternatively, Mathieu et al. (2009) show how the ELA of New Zealand glaciers can be detected using ASTER satellite images. Although considerable variations occur among glaciers in any given year, there is a high degree of reliability for estimating changes across the whole of the Southern Alps when the spatially average ELA_i values are considered. Different types of measurements (e.g. remotely sensed measurements, direct mass balance measurements, geodetic mass balance measurements) each have advantages and disadvantages. It is difficult for direct mass balance measurements from a single glacier to represent entire mountain ranges. Remote sensing measurements allow whole mountain systems to be assessed, but need validation from the mass balance data. Therefore, one future strategy is to blend these different types of measurements, as is done by Huss et al. (2009). Another for New Zealand is to strive

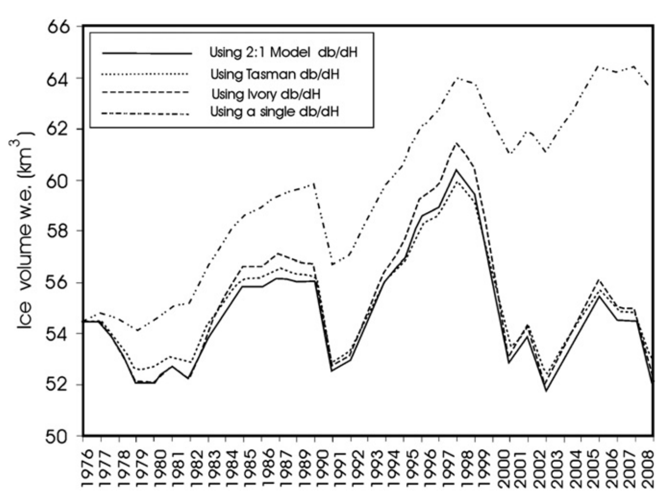

Fig. 12. Sensitivity of ice volume changes to different values of mass balance gradient (*db/dH*). Curves indicate cumulative changes in ice volume (w.e.) for the Southern Alps. The solid line shows model results of this study using mass balance gradients as in Method I (losses from downwasting and pro-glacial lake growth are excluded). These are compared with ice volume changes using: a minimum case ('dry' accumulation and ablation mass balance gradients typical of Tasman Glacier); a maximum case ('wet' accumulation and ablation mass balance gradients typical of Ivory Glacier); and a more simple approach (uses an average single mass balance gradient across the Alps and for both accumulation and ablation areas).

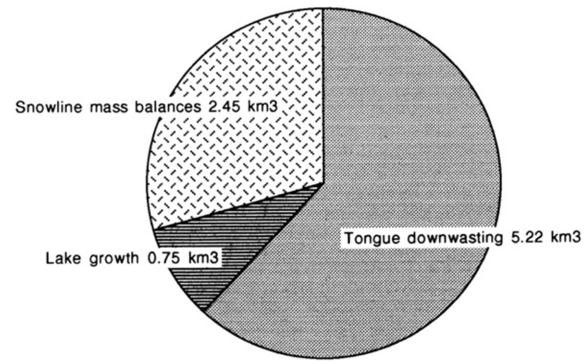

Fig. 13. Sources of ice volume lost from the Southern Alps over the 1976 to 2008 period (total 8.4 km^3 w.e.).

for more long-term glacier mass balance measurements such as initiated in 2004 for Brewster Glacier.

During the IHD, the Ivory Glacier, (area 0.8 km^2) was chosen for intensive measurement (Anderton and Chinn, 1978). These data provide three independent measurements of ice loss (in w.e.):

(1) Repeat mapping using controlled vertical aerial photography, in April 1971 and May 1975. Differences between these two surveys provide a measured 4-year volume loss of -11.78×10^6 m^3.
(2) Field measurements of annual mass balance for the same 4-year period provide a calculated volume loss of -8.44×10^6 m^3.
(3) Using Method I (Table 3) for 1971–1975 provides an estimated ice volume loss of 9.78×10^6 m^3 w.e. This result is within the range provided by the above two independent measurement methods.

The discrepancies among these three estimates for the Ivory Glacier example are attributed to a variety of factors. The Ivory Glacier was far from equilibrium with its climate when the mass balance studies were made. The stake measurements of the mass balance surveys do not take account of ice lost from calving into a small but enlarging pro-glacial lake, and the 'draw-down' effect of increasing ice discharge into the lake. The estimates using aerial photography assume a firn/ice density of 850 kg m^{-3}, a value that is possibly too large for this glacier. In summary, Method I provides estimates of ice loss that appear realistic and within the range of estimated ice volume changes that are obtained independently from photogrammetry and field measurements.

Rapid disintegration of the 12 large protracted response glaciers is a delayed response to disequilibrium with the current climate conditions. They are still responding to climatic warming trend of the 20th century (Chinn, 1996) and are behaving in similar fashion to many other large glaciers elsewhere in the world. For example, ice margin calving into pro-glacial lakes has led to substantial ice losses over the last half century in mountains of Europe (Diolaiuti et al., 2005, 2006), North America (Motyka et al., 2003), South America (Harrison and Winchester, 2000; Naruse and Skvarca., 2000; Raymond et al., 2005) and Asia (Chikita et al., 1999; Sakai et al., 2000; Xie et al., 2007). Tasman glacier is New Zealand's largest and an estimated 135 million m^3 a^{-1} of water is lost from its lowest 10 km (Purdie and Fitzharris, 1999).

The two methods employed in this study require numerous assumptions, so assessment of errors is not straightforward. There are errors associated with the EOSS survey. The length of the ablation season can vary considerably, but for practical reasons, the survey is carried out in March. In years when the snowline on the glacier is obscured, snow patches are used to estimate the altitude of the snow line. From that the equilibrium line altitude is derived. Details of these procedures and likely errors are discussed in detail in Chinn et al. (2005c).

Values for the AAR_0 are assumed to be reasonable, being derived from regression plots of annual *ELA* values for each index glacier against the annual means for the Southern Alps over 32 years. The glacier areas are drawn to World Glacier Inventory standards and have the same accuracy

Table 3
Changes in ice volume of Ivory Glacier as calculated using field measurements of mass balance from stakes, and as calculated using Method I (ELA_0 = 1510 m, minimum elevation = 1389 m, maximum elevation = 1850 m).

Balance year	*ELA* (m)	Glacier area (ha)	Accum. area (ha)	Mean specific b by stake measurements (m w.e.)	Stake volumetric mass balance, *b* × area (10^6 m^3 w.e.)	Method I mean specific *b* (m w.e.)	Method I volumetric mass balance, *b* × area (10^6 m^3 w.e.)
1969–70	1640	79.81	15.57	−2.11	−1.684	−2.34	−1.869
1970–71	1610	79.81	24.71	−1.32	−1.053	−1.49	−1.188
1971–72	1645	74.67	14.69	−1.66	−1.240	−2.39	−1.787
1972–73	1580	69.54	29.63	−1.73	−1.203	−0.67	−0.467
1973–74	1695	64.39	9.09	−3.48	−2.241	−3.25	−2.092
1974–75	1850	59.26	0.00	−4.00	−2.370	−5.95	−3.524
Mean mass balance (m)				−2.38		−2.68	
Total ice volume loss (10^6 m3) April 1969 to May 1975					−9.791		−10.927
April 1971 to May 1975					−7.054		−7.870

as other international glacier inventory studies (e.g. Hoelzle et al., 2007). They have been redrawn from recent surveys to accommodate any glacier recession or lake expansion. Fig. 10 illustrates the 95% confidence limits for estimates of ELA_{Alps} and specific mass balance. For ELA_{Alps}, errors range from 40 to 140 m (from the 20 m contour interval of the 1:50,000 maps) and for specific mass from 0.2 to 2.0 m. These correspond to uncertainty in annual changes of ice volume of 0.23–1.15 km^3 w.e.

Other errors are likely from:

(1) Estimates of mass balance gradients. All available measured values are used here. However, results are sensitive to this term, as demonstrated in Fig. 12. Applying lowest mass balance gradients (no differentiation between "wetter" and "drier" index glaciers) as from Tasman glacier results in a final ice volume change that is 26% less than the study result. Likewise, application of highest mass balance gradients as from Ivory glacier gives an ice volume change that is 11% less. Using an average single mass balance gradient for both accumulation and ablation areas of 10 mm/m results in an ice volume increase of 17% (Fig. 12).
(2) Faulty estimates of pro-glacial lake volumes are possible because of insufficient bathymetric measurements. Bathymetric data has been used for the pro-glacial lake at Tasman glacier terminus from 2003 (Röhl, 2006) and at Godley glacier from 1994 (Warren and Kirkbride, 1998). Volume errors from this source are likely to be less than +/− 10%.
(3) Not applying inherited disequilibrium to those longer response glaciers without pro-glacial lakes. Volume errors from this source are estimated to be less than +/− 5%.
(4) Errors in ELA_{Alps} associated with choice of sample of index glaciers. These are likely to be small (less than +/− 10%) because the index glaciers are derived from a series of transects across all of the Southern Alps and include a wide variety of glaciers.
(5) No snowline flight was undertaken in 1991, but ELA_{Alps} was observed on the Tasman glacier. For this year ELA_{Alps} is calculated using the highly significant linear correlation with ELA_i for the Tasman glacier over 28 years (r = 0.90).
(6) ELA_{Alps} estimates in the 1990 mass balance year are based on reliable data for just two glaciers (Tasman and Marmaduke Dixon). However, these estimates are considered sound because of a highly significant linear correlations (r ~ 0.85) with annual ELA_i for nearby index glaciers.
(7) Errors in ELA_i affecting estimates of annual ice volume change. The mean 95% confidence limits for the annual deviations of ELA_i from ELA_{Alps} are +/− 5 m. This equates to uncertainty in ice volume change of +/− 0.6 km^3, or +/− 7%.

Most of these error sources are compensatory rather than additive and estimates of ice volume changes are derived from actual and widespread measurements made over many years. Errors are likely to be smaller when compared with parameterized studies. For example, the largest source of volume loss for the whole Southern Alps is from downwasting of the 12 large protracted response glaciers and this is measured directly by detailed aerial mapping and ground surveys. The collective assessment of the above error sources suggests that the overall error range for ice volume changes is within +/− 20%.

Previous studies have shown that overall annual behaviour of ELA_i for the index glaciers is typical of that for the whole Southern Alps (Lamont et al., 1999; Clare et al., 2002). Thus the Southern Alps behave as a unified climatic unit, despite being crossed by a steep precipitation gradient. Changes in annual ELA_{Alps}, inferred mass balances and calculated changes in ice volume for the period 1977–2008 are consistent with year to year variations in climate parameters, airflow patterns over the Southern Alps and wider atmospheric circulation patterns as reported by others (Fitzharris et al., 1992, 1997; Tyson et al., 1997; Lamont et al., 1999; Clare et al., 2002; Fitzharris et al., 2007) and ties into previous work on synoptic weather and climate patterns (Kidson, 2000; Renwick, 2011).

Annual temperature, driven by circulation patterns, appears to be much more important than precipitation. Years with positive mass balances for the Southern Alps are characterised by below average temperature anomalies, enhanced southwest flow and more troughs over New Zealand. Years with negative mass balance have above average temperature, reduced southwest flow and more blocking, especially east of New Zealand. For much of the century before the survey period begun in 1976, most glaciers underwent wholesale recession and loss of ice volume. Ice volume at about 1850 is estimated at 170 km^3 (Hoelzle et al., 2007), compared with about 100 km^3 in 1880 (Ruddell, 1995), 55 km^3 in 1976 (Chinn, 2001) and 46 km^3 in 2008. The rate of ice loss between 1976 and 2008 equates to 0.3 $km^3\,a^{-1}$. However, this is probably slower than the rate of ice loss between the 19th century and 1977, which is estimated to have varied between 0.5 $km^3 a^{-1}$ and 0.8 $km^3 a^{-1}$. With hindsight it appears that the ELA surveys began about the time of a fundamental switch in atmospheric regime over New Zealand associated with changes in the IPO (Salinger et al., 2001). Salinger and Mullan (1999) note a period of anomalous east and north east airflows over New Zealand up until 1975, followed by anomalous west to south west airflows from 1976 onwards. These circulation changes may have favoured a switch to lower snowlines and more frequent occurrence of positive mass balances. As a result, widespread glacier recession of small and medium glaciers was halted and many advanced or thickened in their upper reaches. These observations agree with the conclusion of Anderson and MacKintosh (2006) that temperature change is the major driver of Holocene glacier fluctuations in New Zealand.

Comparisons of climatic conditions with specific mass balance show that higher ELA_{Alps} and negative mass balance all occur in glacier years with temperatures above the 1971–2000 average. Six of these years (1978/79, 1989/90, 1999/2000, 2001/02, 2005/06, and 2007/08) had below average regional precipitation, more anticyclones and anomalous easterly circulation. For 1989/90, temperatures were +0.5 °C above the average. Another year (1998/99) was characterised by anomalous northwesterly circulation, which brought above average precipitation to the Southern Alps, but much warmer temperatures (+0.7 °C). Lower ELA_{Alps}

and positive mass balance years normally occur when cooler westerly and southwesterly circulation anomalies dominate during summer. Typically these circulation patterns deliver above average precipitation for the Southern Alps, occur with increased frequency during El Niño years, and characterised the glacier years 1982/83–1984/85, 1991/92, 1992/93, 1996/97 and 2004/05.

7. Conclusions

Traditionally, mass balance measurements have been used to estimate changes in glacier volume, usually measured in water equivalent. Without doubt, they provide high quality data that is of great scientific value. However, this approach is very resource intensive and not always suited for sparsely population regions with limited funding resources. One need is to find ways of estimating mass balance and glacier volume changes for whole mountain systems. A second need is to find more cost effective, indirect ways of assessing glacier volume changes for many parts of the world. The New Zealand record of annual snowline measurements for the Southern Alps since 1977 (Chinn et al., 2005c) is used here as a case study as to how these needs might be met.

Based on annual end-of-summer snowline data from the 50 index glaciers, this paper presents a practical method to estimate annual changes of glacier ice volume for a substantial mid-latitude mountain system, the Southern Alps of New Zealand. Measurements of snowlines are used to calculate annual mass balance using a variation of the method of Rabatel et al. (2005). A critical parameter is the vertical gradient of mass balance. Various values are used—one for the accumulation area, and the other for the ablation area. Different values are also chosen for the wetter area west of the Main Divide and for drier areas to the east. With this information and the New Zealand glacier inventory, ice volume changes have been calculated for all but the 12 large protracted response glaciers of the Southern Alps for the period 1976–2008. When this method is applied to the Ivory Glacier, there is good agreement (within 15–20%) with changes in ice volume as measured using two independent methods.

Total water equivalent of the ice volume for the Southern Alps of New Zealand was assessed at 54.5 km^3 in 1976. Between 1977 and 2008 this ice volume has been reduced by an estimated 8.4 km^3 (water equivalent). Most glaciers are small to medium in size. Snowline data and estimates of mass balance for these indicate a net loss of 2.45 km^3 water equivalent. This is less than one third of the total ice loss for the Southern Alps. More than two thirds (5.96 km^3) has come from just 12 of New Zealand's large protracted response glaciers and is due to tongue downwasting and terminus calving into expanding pro-glacial lakes. For these, the snowline method is not appropriate and ice loss is estimated from repeat geodetic surveys. Overall, 15% of the total ice volume of the Southern Alps has been lost between 1976 and 2008. This equates to a rate of $-0.3\ km^3\ a^{-1}$, but this is less than half the rate of ice loss estimated for the previous 100 years.

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