THE CONTEMPORARY RETREAT OF TASMAN GLACIER, SOUTHERN ALPS, NEW ZEALAND, AND THE EVOLUTION OF TASMAN PROGLACIAL LAKE SINCE AD 2000

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With 4 figures, 3 tables and 5 photos
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Summary: The Tasman Glacier is one of many freshwater-terminating glaciers located in the *Aoraki*/Mt Cook National Park that has retreated significantly since the mid-20th century. Although there have been several observations made of the glacier since the formation of Tasman Lake and the following rapid retreat, the glacier entered a period of rapid terminus disintegration and retreat since c. AD 2000. The retreat of Tasman Glacier between 2000 and 2008 has occurred in two distinct periods: an initial period of relatively slow retreat prior to 2006, followed by a secondary period of rapid retreat between 2006 and 2008. Terminus full width retreat for the period 2000–2006 occurred at a rate of 54 m a^{-1}, accelerating to 144 m a^{-1} during 2006–2008. During the period 2000–2006, the controlling process of ice loss at the terminus was iceberg calving resulting from thermal undercutting. In contrast, the retreat between 2006 and 2008 was probably controlled by buoyancy-driven iceberg calving caused by decreased overburden pressure as a result of supraglacial pond growth. As a result, the surface area of Tasman Lake has increased by 86% over the period 2000–2008, with lake volume increasing by 284% between 1995 and 2008. Currently, the volume of Tasman Lake is 510×10^6 m^3. It will increase dramatically in near future as the lake expands into deeper water.

Zusammenfassung: Der Tasman Glacier ist einer der zahlreichen in proglaziale Seen kalbenden Gletscher im *Aoraki*/Mt Cook National Park, welche sich seit der Mitte des 20. Jahrhunderts stark zurückgezogen haben. Obwohl mehrere Beobachtungen seit der Bildung des Tasman Lake und nachfolgendem Rückzug vorliegen, befindet sich der Gletscher seit dem Jahr 2000 in einer Periode schneller Auflösung der Gletscherzunge und raschen Rückzugs. Der Rückzug des Tasman Glacier zwischen 2000 und 2008 trat in zwei unterschiedlichen Phasen auf: einer ersten Phase relativ langsamen Rückzugs vor 2006, gefolgt von einer zweiten Phase schnellen Rückzugs zwischen 2006 und 2008. Der gemittelte Rückzug der Gletscherfront von 2000 bis 2006 betrug 54 m a^{-1} und beschleunigte sich auf 144 m a^{-1} von 2006 bis 2008. Während der Phase von 2000 bis 2006 wurde der Eisverlust an der Gletscherfront durch Abkalbung von Eisbergen durch thermales Unterschneiden kontrolliert. Im Gegensatz dazu wurde der Rückzug zwischen 2006 und 2008 wahrscheinlich durch Auftriebsdruck gesteuertes Abkalben von Eisbergen durch Verringerung des auflastenden Drucks infolge des Wachstums von supraglazialen Schmelzwassertümpeln verursacht. Als Ergebnis wuchs die Oberfläche des Tasman Lake während der Periode von 2000 bis 2008 um 86% an, das Seevolumen um 284% zwischen 1995 und 2008. Aktuell beträgt das Volumen des Tasman Lake 510 x 10^6 m^3. Es wird in der näheren Zukunft dramatisch ansteigen, da der See sich zu tieferem Wasser ausdehnt.

Keywords: Calving, Tasman Glacier, recession: 21st century, debris-cover, New Zealand

1 Introduction

Mountain glaciers receive a reputation as high-resolution archives of climate variations (IPCC 2007). Because of their importance as key indicators of past, present, and future climate change, intense efforts have been made to monitor and understand changes of glacier volume and position (Bamber and Payne 2004; Knight 2006; Zemp et al. 2008). However, the insulation of glacier ice under a thick supraglacial debris cover often decouples glaciers from changes in climate compared with similar clean ice glaciers (cf. Kirkbride and Brazier 1998, Purdie and Fitzharris 1999; Nakawo et al. 2000; Winkler 2009). The Southern Alps of New Zealand provides an excellent example of this, with almost debris-free maritime glaciers (Franz Josef and Fox Glaciers) situated on the western flanks of the Alps responding rapidly to changes in climatic input throughout the past century (Chinn et al. 2005; Anderson et al. 2006; Purdie et al. 2008). This has been represented by periods of advance and retreat visible in changes in terminus position over decades.

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Contrary to their western counterparts, the termini of the larger, debris-covered glaciers of the *Aoraki*/Mt Cook region, situated on the eastern flank of the Southern Alps, have shown a pattern of fluctuations less intimately connected to the regional climate trends over the last century (Kirkbride 1993; Chinn 1996; Hoelzle et al. 2007). This difference is primarily due to the low gradient long-profiles and the presence of insulating supraglacial debris-cover on the glaciers of the *Aoraki*/Mt Cook region (Hambrey and Ehrmann 2004), and has resulted in formation of down-wasting glacier termini, allowing development of a number of proglacial lakes (Kirkbride 2000). These lakes are generally constrained by post-Last Glacial Maximum (LGM) outwash gravels, on which Little Ice Age (LIA) and older late Holocene moraines are situated (Gellatly 1985; Röthlisberger 1986; Winkler 2004, 2005; Burrows 2005; Schaefer et al. 2009). The formation of these proglacial lakes has further decoupled these glaciers from changes in climate, with the transition to calving termini appearing to over-ride climatic inputs (Kirkbride and Warren 1999; Warren and Kirkbride 2003). Tasman Glacier, the largest temperate valley glacier in the Southern Alps of New Zealand, is the most famous example of a glacier responding in this fashion.

During the most recent period of retreat (2000–2008), Tasman Glacier has entered a period of rapid terminus disintegration that has further increased the rate of recession. This study seeks to quantify the increase in the retreat rate for Tasman Glacier over the period 2000–2008 through the use of sequential satellite images and bathymetric mapping of Tasman Lake, conducted in April 2008.

2 Study area and local setting

Located east of the Main Divide of the Southern Alps in the *Aoraki*/Mt Cook National Park (Figs. 1, 2), Tasman Glacier is New Zealand's largest temperate valley glacier currently extending ~23 km down the Tasman valley from a névé at 2400 m, to c. 717 m at the terminus (Photo 1). The glacier covers an area of c. 98 km^2 (total area according to the World Glacier Inventory in 1978, tributary glaciers not included), and varies between 1 and 2 km in width (Hochstein et al. 1995). The Tasman Glacier terminus currently receives ice from a number of tributary glaciers including the Ball, Hochstetter, Haast, and Rudolf glaciers (Fig. 1). However, the primary contemporary source of ice is currently the Hochstetter and Ball Glaciers,

Photo 1: Oblique air photo of Tasman Lake and the lower glacier tongue of Tasman Glacier (Photo: S. Winkler, March 2008)

indicated by the decline in surface ice velocity immediately up-glacier of the Tasman-Hochstetter Glacier confluence, followed by a dramatic increase in ice velocity down-glacier of this confluence (Anderton 1975; Kääb 2002). Summer surface ice velocities measured by Röhl (2008) in 2001/02 decreased dramatically from 96 m a^{-1} near Ball Glacier to less than 3 m a^{-1} at the Tasman Glacier terminus.

Down-valley of the Tasman-Hochstetter Glacier confluence, a large supraglacial debris cover has formed from both low-magnitude/high-frequency rock falls and rock avalanches (Photo 2; Kirkbride and Warren 1999). The debris cover consists of very poorly sorted, predominantly clast-supported angular boulders of sandstone and argillite sourced from the surrounding slopes (Kirkbride and Warren 1999). Lateral spreading of the supraglacial material down-valley of the confluence produces a continuous debris cover across the glacier surface. A larger accumulation of debris (~ 1 m thick) occurs at the glacier terminus (Kirkbride 1989; Röhl 2008), although several narrow tongues of clean ice have been observed extending down-glacier (Fig. 1 of Kirkbride 1993; Kirkbride and Warren 1999).

The extensive debris-cover present on the lower reaches of the Tasman Glacier has meant that the 20th century history of the glacier has been predominantly one of progressive thinning with no change in terminus position. In contrast to its nearest neighbouring glaciers (Mueller and Hooker Glaciers), Tasman Glacier was still at its LIA-maximum position around 1900 with historical evidence of an "overtopping" of the existing LIA-moraine as late as c. 1910 (Burrows 1973; Gellatly 1982; Winkler 2004). This would give a delay in timing of the LIA-maximum of at least 150 years and demonstrates the unique case of this glacier.

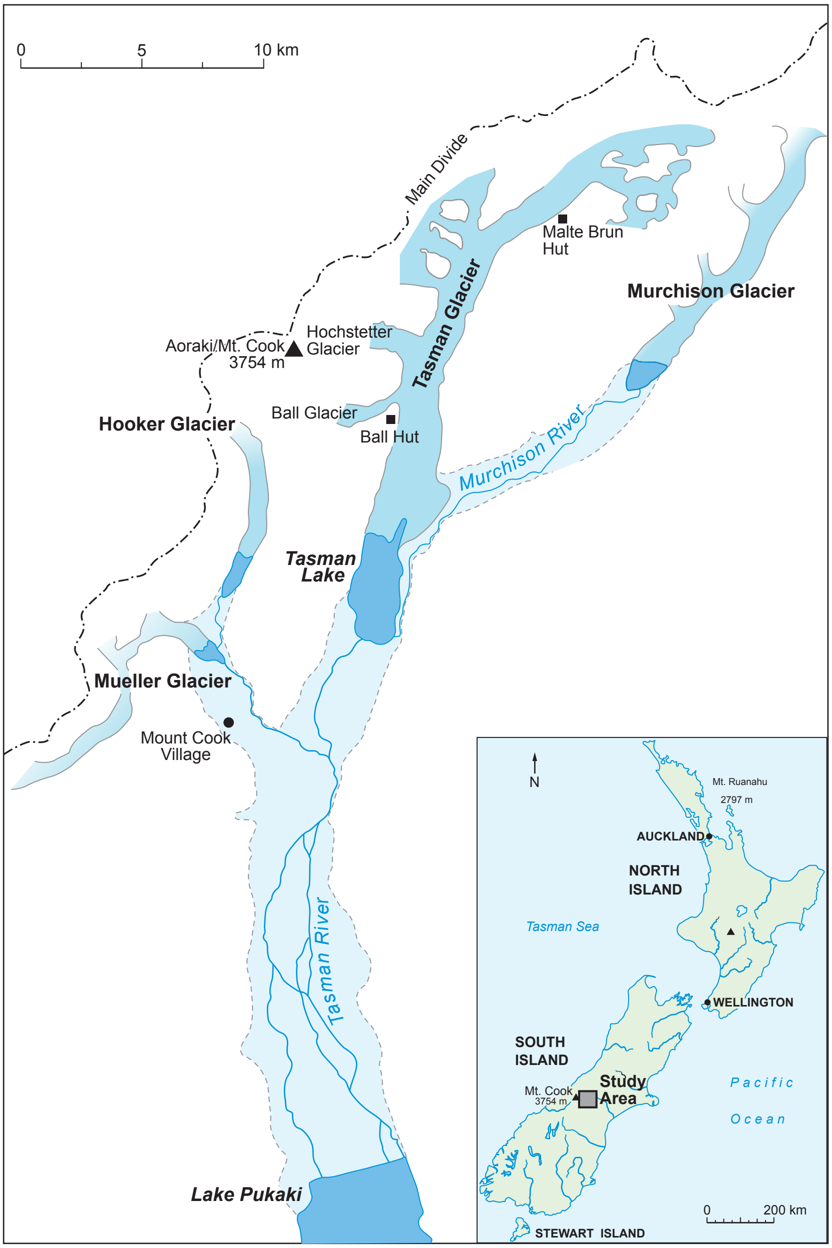

Fig. 1: Location map of the study area and Tasman Glacier

The debris cover on Tasman Glacier has accumulated from a combination of meltout of englacial material (Kirkbride and Spedding 1996), rock avalanching directly onto the glacier surface (Sugden and Kirkbride 1992) and collapse of lateral moraine walls (Blair 1994). As an averaged negative mass balance was maintained during most of the 20th century following the LIA, the lower glacier tongue experienced a mass loss as the mass transfer was not adequate to compensate ablation. This meant that the Tasman Glacier terminus lowered to a gradient of less than 2°, providing the preconditions for supraglacial pond development (Reynolds 2000). Large amounts of sediment have also been supplied to the proglacial zone during extended periods of glacier lowering, allowing formation of large laterofrontal moraines by dumping processes, and of an outwash head (Kirkbride 1993, 2000).

As down-wasting proceeded, perched meltwater ponds developed supraglacially, sourced from the

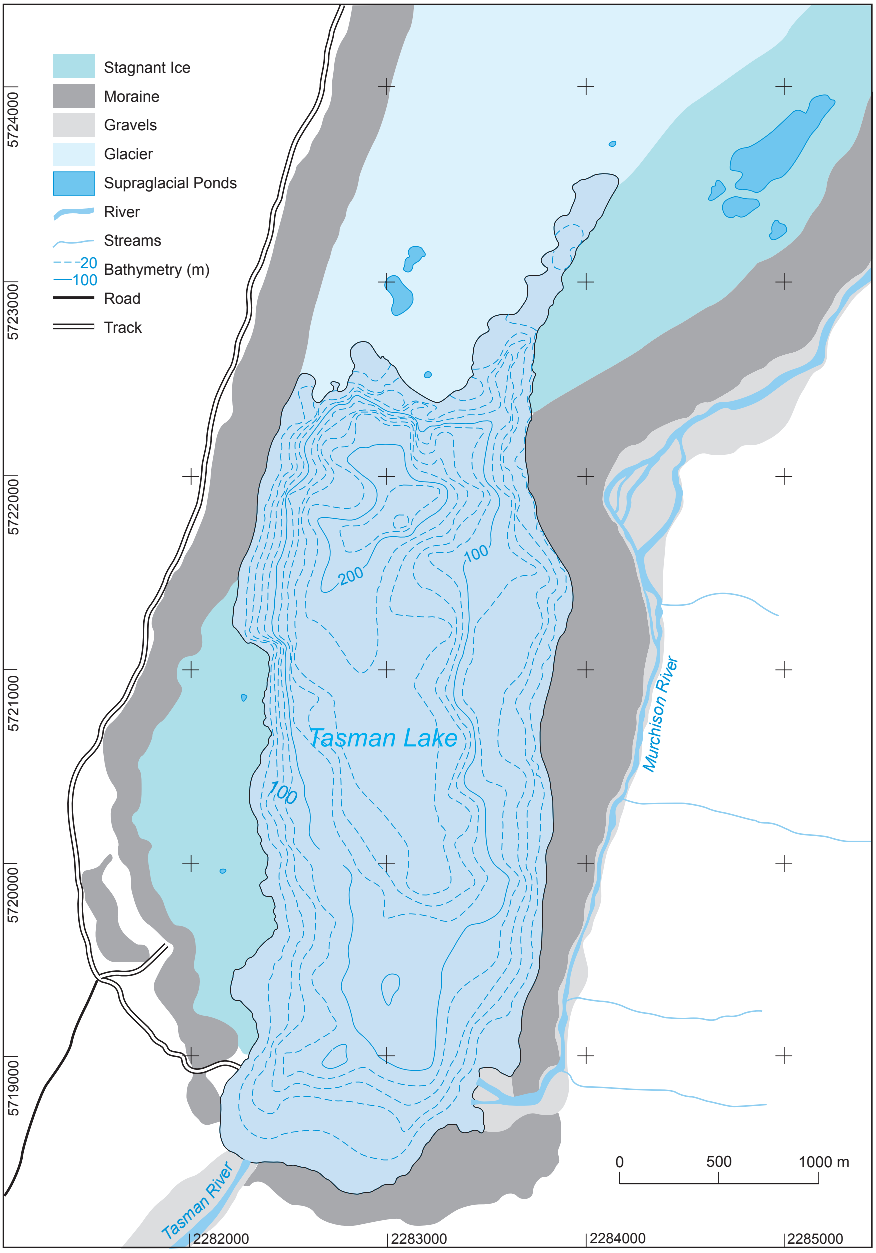

Fig. 2: Detailed map of the Tasman Lake showing lake bathymetry relative to lake surface at 717 m a.s.l.

melting of clean ice exposed due to the redistribution of supraglacial debris during surface lowering. Drainage eventually penetrated downwards, intercepting englacial conduits, to develop hydraulically connected supraglacial lakes (Kirkbride 1993). At this point, supraglacial lakes reflected the local piezometric surface within the glacier. Englacial conduits and crevasses representing lines of weakness within the glacier provided the loci for lake development from this point onwards. Larger supraglacial lakes were formed during the mid-1980's along these lines of weakness which were either solely ice-based, or penetrated to the subglacial bed (Kirkbride 1993).

Adjacent supraglacial ponds increased in size during the mid-1980s as a result of increased melt of clean ice and iceberg calving at lake margins, eventually coalescing to form a single lake (Tasman Lake) by 1991 (Kirkbride 1993). As recently as this point, Tasman Glacier began to retreat from its

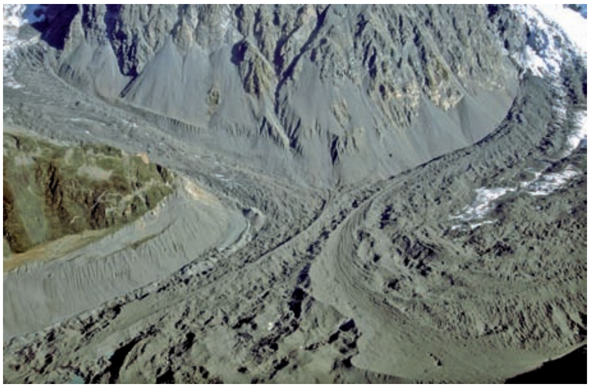

Photo 2: Oblique air photo of the supraglacial debris cover on Tasman Glacier. The confluence of Ball Glacier can be seen on the left, the confluence of Hochstetter Glacier to the right (Photo: S. Winkler, March 2006)

LIA terminus position. The rapid retreat of Tasman Glacier since then has been controlled by the complex interaction between glacier dynamics, surface slope and geometry, and the presence of Tasman Lake (Kirkbride and Warren 1999). Tasman Glacier has one of the longest and most detailed histories of all New Zealand glaciers (Tab. 1; cf. detailed reviews in Gellatly 1985; Kirkbride and Warren 1999; Burrows 2005). Originally surveyed by von Haast (1866) and von Lendenfeld (1884), the Tasman Glacier has been studied intermittently throughout the 20th century. The studies that have been undertaken at Tasman Glacier have ranged from mapping of the glacier surface (Kirkbride 1989), geophysical cross-profiling of the glacier bed at Malte Brun and Ball huts (Broadbent 1973; Anderton 1975; Claridge 1983; Watson 1995), measurement of ablation rates for clean ice and debris cover (Kirkbride 1989; Purdie and Fitzharris 1999), and most recently, bathymetric mapping of the incipient Tasman Lake (Hochstein et al. 1995; Watson 1995; Röhl 2005).

3 Methodology

3.1 Terminus position and lake area

To acquire the historical Tasman Glacier terminus position and Tasman Lake area, 12 multispectral satellite images were collected for the period of 29/04/00 to 18/10/08 from which terminus positions could be delineated. The satellite images used in this study were obtained from three different satellite sources, the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER), Systeme Pour l'Observation de la Terre (SPOT), and Quickbird. The satellite images were geo-referenced to the highest resolution image available.

Geo-referencing was undertaken using ENVI 4.5© image-to-image registration method, where ground control points (GCPs) are used to register a selected image to a base image. In this study the 23/01/2007 SPOT image geo-referenced to New Zealand Map Grid (NZMG 1949), was used as the base image from which all other images were registered, using conspicuous and stable features present in every image. All other satellite images were registered using the rotation, sampling and translation (RST) warping method (e.g. Jensen 2007), fixed to over 20 GCPs. The resulting root mean square error (RMS) for each rectified image was smaller than a single pixel. The resulting accuracy of the digitised terminus positions from the satellite images is estimated at ~15 m (Röhl 2008).

Retreat rates in this study were calculated by measuring the perpendicular distance between terminus profiles (Fig. 3) at a series of locations along the ice front. The change in distance for each perpendicular measure was annually averaged, with the mean rate of change taken from this information to represent the rate of change during the study period. From the digitised terminus positions retreat rates could be calculated for the Tasman Glacier during 2000–2008. Nine periods were used to calculate retreat: 29/04/00–07/04/01, 07/04/01–14/02/02, 14/02/02–31/12/02, 31/12/02–09/09/05, 09/09/05–29/04/06, 29/04/06–23/01/07, 23/01/07–08/04/07, 08/04/07–13/05/08, and 13/05/08–18/10/08.

3.2 Lake bathymetry

The mapping of Tasman Lake bathymetry was carried out in April 2008 through the integration of bathymetric and global positioning system (GPS) measurements operated from an inflatable boat. A Humminbird® 323 DualBeam Plus™ dual frequency echo-sounder using operating frequencies of 200 kHz and 83 kHz and a hand-held GPS unit were used to obtain non-continuous water-depths throughout Tasman Lake. The accuracy of water-depth (z) measurements for this study were not recorded, however, Röhl (2005) reported an accuracy of ± 1 m over a range of 5 to 125 m when using a Humminbird 200 kHz single beam echo-sounder. Point position (x, y) accuracy for the GPS units is ± 5 m.

Tab. 1: Observations and studies of the Tasman Glacier terminus and Tasman Lake (adapted and updated after Watson 1995)

Year	Surveyor	Comments
1866	Von Haast	Plane table survey of the Tasman Glacier terminus
1882	Green	Ice from Ball Glacier overtopped lateral moraines at Ball Hut corner
1884	Von Lendenfeld	Survey of Tasman Glacier. 1:80 000 map produced position of terminus slightly further back than 1862 map.
1889–1890	Harper	Ice level at Ball Glacier junction still above lateral moraine.
1891 and 1906	Broderick	Terminus surveyed showing minor advance between 15 and 45 m between the two surveys
1905–1972	Harper	Terminus position remains largely unchanged. 21 m of downwasting at Ball Hut corner between 1890 and 1924.
1930s	Harper / Rose	Down-wasting probably did not begin at the terminus until 1913/14.
1960s	McKellar	No significant change in position of the terminus.
1966–1970	Anderton (1975) Ministry of Works, DSIR	Glaciological and Geophysical studies from 1966 to 1973 give approximate ice thickness and specific balances at Tasman Saddle, Malte Brun and Ball Huts.
1972	Broadbent (1973) Ministry of Works, DSIR	Seismic and gravity surveys on the Tasman Glacier indicating a maximum glacier thickness of c. 615 m at Ball Hutt and c. 600 m and Malte Brun Hut.
Late 1960s	Hochstein	Rapid melting began in a few isolated melt ponds in the centre of the terminus region and in a small elongated lakelet towards the eastern moraine.
1982	Claridge (1983)	Geophysical study of the Tasman Glacier terminus. The surface of the terminus lowered at a rate of 2.0 – 3.4 m a^{-1} between 1972 and 1982.
1986	Kirkbride (1989)	Debris-cover has significantly affected the flow structure of the glacier. Twentieth-century negative mass balance has resulted in the formation of thermokarst lakes at the terminus.
Late 1980s	Hochstein	Melt pools coalesce to form a large melt lake.
1993	Hochstein et al. (1995)	Bathymetric survey of Tasman Lake indicating a surface area of ~2 km^2.
1994	Watson (1995)	Measurements at Ball Hut and Upper Tasman show downwasting of 2-3 m a^{-1} and 0.8 m a^{-1} respectively.
1995	Watson (1995)	Continued expansion of Tasman Lake, now with a surface area of ~2.35 km^2.
1999	Purdie and Fitzharris (1999)	Ablation rate at Tasman Glacier terminus calculated to be 7 mm d^{-1} under debris-cover, 96 mm d^{-1} for clean ice.
1999	Röhl (1999)	Study of Tasman Lake sedimentation indicating a high degree of spatial variation in sedimentation rates.
2002–2005	Röhl (2005)	Tasman Lake enlarges to a surface area of 3.7×10^6 m^2 as the Tasman Glacier retreats up-valley due to iceberg calving at the glacier terminus.

As the echo-sounder used in this study reports water depth at the point of greatest reflection of the pulse emitted, difficulties were encountered close to the glacier's terminal ice cliff, due to the presence of a submerged ice "foot" sloping away from the ice cliff. As a result, water depths at the ice cliff were not possible to obtain, with constant depths only recorded at a point away from the ice cliff (typically within 50 m). Bathymetric data was then interpolated using triangular irregular network (TIN) modelling, from which smoothed bathymetric contours were produced (Fig. 2).

4 Results

4.1 Lake area and volume

The increases in Tasman Lake surface area and volume are shown in table 2 and figure 4. As is evident from figure 4, the rate at which Tasman Lake has increased in size since the mid-1950s has been dramatic. Since the coalescing of supraglacial ponds in the mid-1980s the rate at which Tasman Lake has expanded has increased rapidly. Surface area expan-

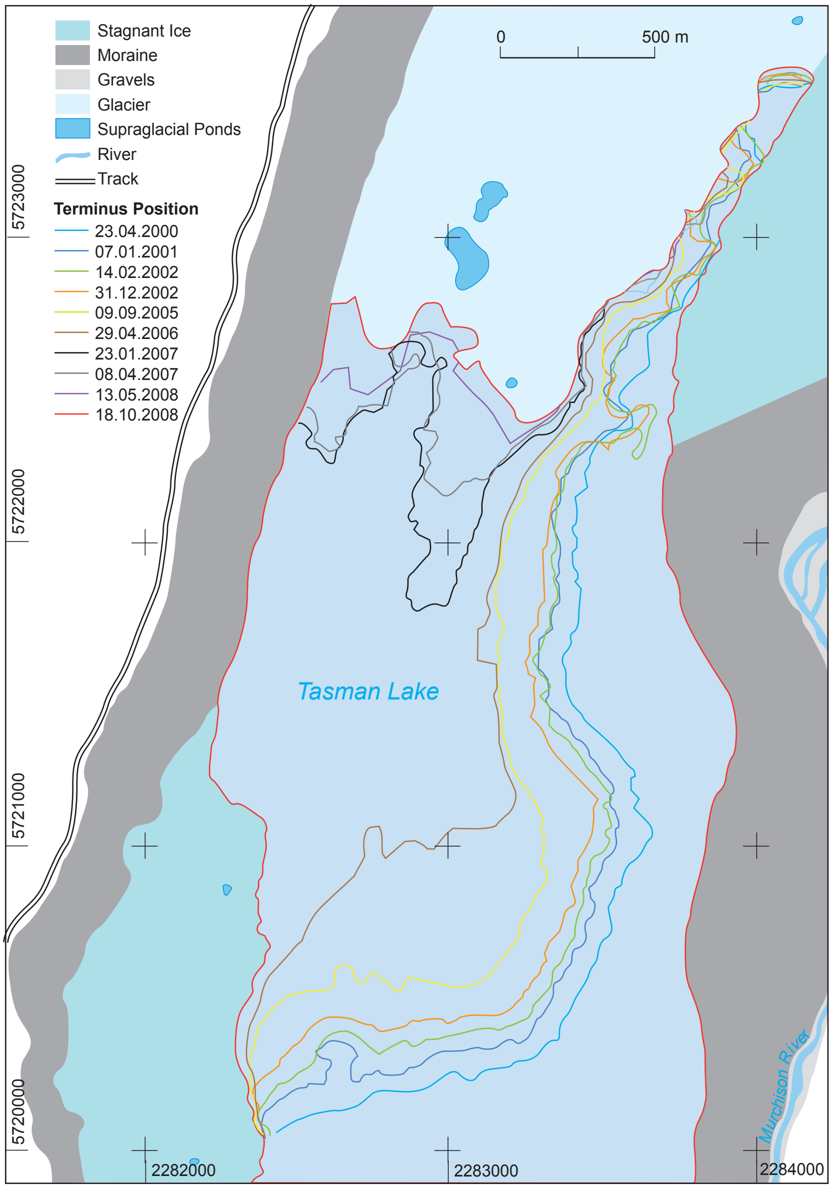

Fig. 3: Terminus position of the Tasman Glacier for the period of 23/04/2000 to 18/10/2008, derived from sequential satellite imagery

sion of the lake occurred at a mean rate of 0.14 × 10^6 $m^2 a^{-1}$ between 1982 and 1995. Concomitant increases in lake volume have also occurred during this period of lake growth, with the lake increasing in volume to 133 × 10^6 m^3 over the period 1982 to 1995 (Tab. 2).

Recent changes occurring at the terminus of Tasman Glacier have resulted in the further expansion of Tasman Lake up-valley. Indeed, the surface area of Tasman Lake has increased significantly over the period 2000–2008 (Fig. 4), with the lake expanding at a mean rate of 0.34 × 10^6 $m^2 a^{-1}$ for this period (Tab. 2). This rate of surface area increase is significantly higher than previously recorded period of lake growth between 1982 and 1995.

The volume of water held by Tasman Lake has also increased considerably as Tasman Lake has expanded (Photo 3). The increase in lake volume is a function of not only the increase in surface area of Tasman Lake, but also due to Tasman Glacier retreating into a significantly deeper part of the glacially-excavated trough (Fig. 2). This has resulted in the lake attaining a net volume of 510 ×10^6 m^3 in 2008, which represents a volume increase of 284% (377 × 10^6 m^3) between 1995 and 2008.

4.2 Lake bathymetry

Tasman Lake bathymetry (2008) is shown in figure 2. The lake as a whole has a relatively uniform floor, with little irregularity especially throughout the southern end of the lake. Up-valley towards the glacier terminus, Tasman Lake increases in depth, with the lake floor becoming more irregular and complex in morphology. The lake deepens gradually up-valley of the LIA moraines until it increases in depth dramatically c. 800 m from the current glacier terminus. From this point onwards, the lake shallows rapidly towards the glacier front indicating the presence of an ice foot. However, the bathymetric slope is not uniform, highlighting the spatial variability and complexity of processes acting at the glacier terminus.

The current maximum lake depth of 240 m is attained 800 m in front of the centre of the glacier terminus (Fig. 2), located within a trough that runs along the centre-right of the lake. In general the lake floor follows the typical pattern of glacial valley cross-profile morphology (Brook et al. 2006), though the lake floor rapidly slopes up toward the surface of the lake on the true right, with the true left of the lake floor having a more gently sloping form.

4.3 Terminus position

Overall, retreat of Tasman Glacier has been rapid (Fig. 3), although significant temporal variations in the rate of retreat (u_r) of the terminus can be identified during the study period (2000–2008). Specifically, there is an identifiable transition in the rate of retreat prior to, and after, 2006. For the period 2000–2006 the glacier retreated a mean distance (Δx_{mean}) of 215 m, equating to a u_r of 54 m a^{-1} (Tab. 3). However, the actual annual frontal retreat rate was highly varied *within* each period pre- and post-2006. As a result, u_r has ranged between 33 and 145 m a^{-1}, indicating a temporal variability in the process-mechanisms operating along the Tasman Glacier terminal ice cliff.

The rate of retreat of the Tasman Glacier terminus increased dramatically between 2006 and 2008, signifying an even faster rate of retreat. During this time, the glacier's ice cliff retreated a mean distance (Δx_{mean}) of 357 m, resulting in a u_r of 144 m a^{-1} (Tab. 3). This rate of retreat is significantly higher compared to the previous period of retreat documented by this study. However, as with the previous period of retreat (2000–2006), the retreat rates calculated between 2006 and 2008 have varied significantly, with u_r ranging between 123 and 258 m a^{-1} (Tab. 3). Furthermore, maximum terminus retreat (Δx_{max}) during this period occurred during a 9 month pe-

Tab. 2: Surface area, volume and lake level of Tasman Lake between 1957 and 2008

Date of Survey	Surface Area ($\times 10^6$ m^2)	Water Volume ($\times 10^6$ m^3)	Source
1957	0.01	-	Kirkbride (1989)
25/02/1964	0.06	-	Kirkbride (1989)
01/09/1972	0.08	0.00	Hochstein et al. (1995)
1982	0.56	11.00	Hochstein et al. (1995)
01/02/1986	1.07	-	Kirkbride (1989)
30/04/1993	1.95	125.00	Hochstein et al. (1995)
01/02/1995	2.35	133.00	Watson (1995)
29/04/2000	3.21	-	*this study*
07/04/2001	3.48	-	*this study*
14/02/2002	3.67	-	*this study*
31/12/2002	3.83	-	*this study*
09/09/2005	4.70	-	*this study*
09/02/2006	4.92	-	*this study*
23/01/2007	4.96	-	*this study*
08/04/2007	5.78	-	*this study*
13/05/2008	5.96	510.00	*this study*

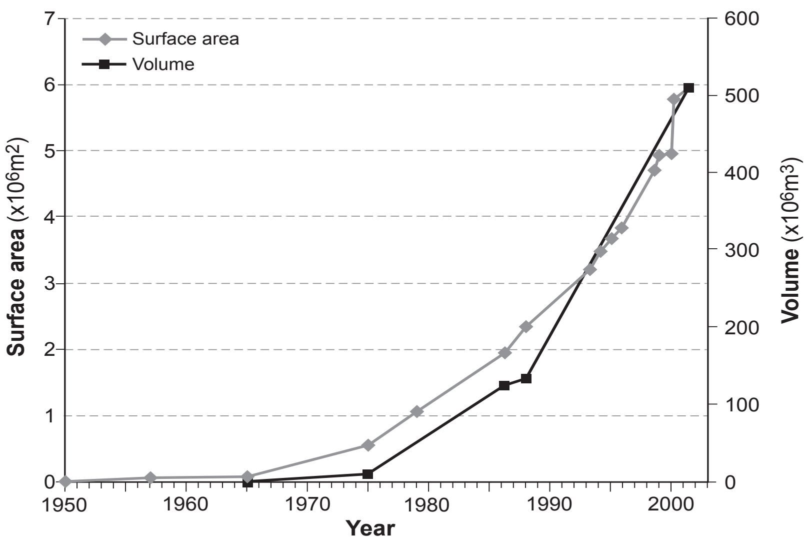

Fig. 4: Surface area and water volume of meltwater ponds (pre-1980s) and Tasman Lake between 1957 and 2008, excluding a c. 10 m drop in lake level in 1995 (cf. Tab. 3)

riod between 29/04/2006 to 23/01/2007, when the true right of the glacier retreated c. 1600 m. This equates to a mean retreat of 5 m d^{-1}, though in reality, the terminus has retreated non-linearly over these shorter timescales (Fig. 3).

5 Discussion

5.1 Tasman Lake

After an initial modest rate of lake volume increase during the early stages of lake development, Tasman Lake has proceeded to increase in surface area and volume at a rapid rate (Fig. 4, Tab. 2), in line with the stages of development outlined by Kirkbride (1993). As the lake has expanded due to the retreat of Tasman Glacier, significant changes in lake bathymetry have occurred.

Differences in lake floor bathymetry occur between the highly irregular northern end of Tasman Lake, at the present terminus, and the relatively uniform southern end of the lake, at the outlet of the Tasman River (Fig. 2). This indicates that the lake floor directly in front of the terminus at the north end of the lake may be largely ice-cored. This may in fact signify that rather than a low-angle "ice foot" along the full length of the present terminus, partial or full depth subaqueous calving occurs, with an ice foot present only along the embayment running along the true left of the terminus. At the centre of the current terminus the maximum lake depth of 240 m is attained, indicating that retreat of the total depth of the glacier is now proceeding, with fracturing and flotation of the "ice foot" having occurred recently.

Ice-cored areas of lake floor at the terminus are the result of subaqueous calving processes acting at depth. Supraglacial material sliding off the terminal cliff, meltout of englacial debris and the settling of suspended sediment from subglacial meltwater (Drewry 1986) has led to the accumulation of deep lake floor sediments. As the ice-cored sections of the lake floor melt and debris cover becomes redistributed, buoyancy forces will act to detach these sections of ice from the lake floor (Röhl 2005), producing subaqueous kettle holes (Gustavson 1975) that represent the greatest depths in the northern part of the lake.

5.2 Terminus position

The width-averaged annual rate of retreat (u_r) calculated for the Tasman Glacier over the period 2000–2006 (Tab. 3) occurred at a similar rate to initial studies of Tasman Glacier retreat since Tasman Lake began forming. Indeed, the earliest reported rate of retreat is from Warren and Kirkbride (2003), who identified a retreat rate of 34 m a^{-1} over the period of 1994–1995. Retreat rates for their study were derived from ground surveys and near-vertical and oblique aerial photographs, with vertical aerial photographs from 1958, 1965, 1974 and 1985/86 used to indicate historical patterns of frontal changes. Compared to the present study, the rate of retreat reported by

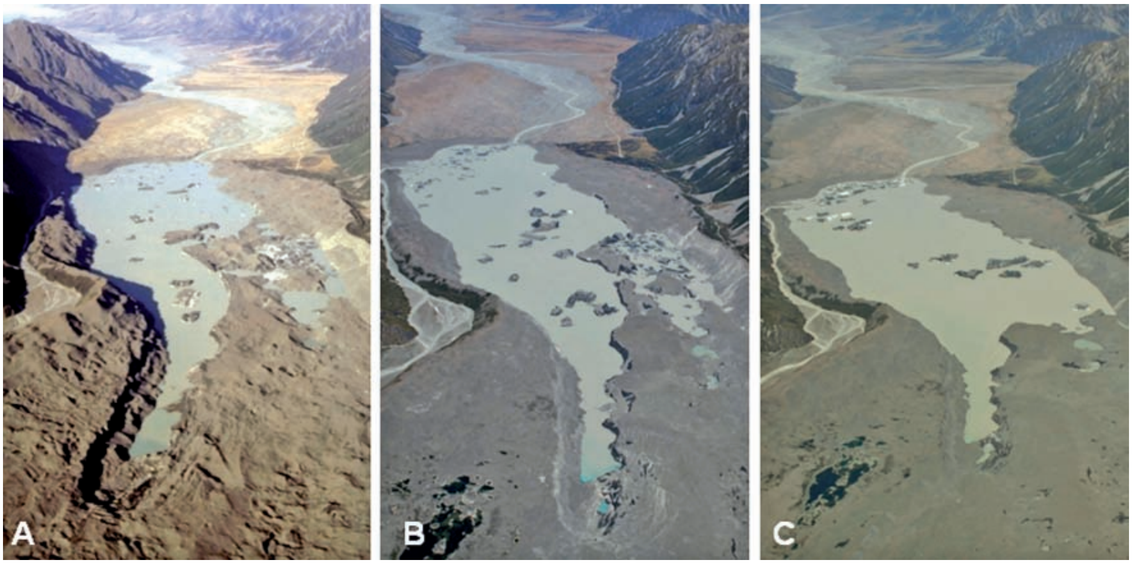

Photo 3: Oblique aerial photographs showing the disintegration of the lower Tasman Glacier terminus between (A) March 2006, (B) March 2007, and (C) March 2008 (Photos: S. Winkler)

Warren and Kirkbride (2003) is rather low, but still within the annual variation indicated in table 3.

There is also a congruence between results of this study and that of Röhl (2006), who identified a retreat rate of 54 m a^{-1} (2000–2003) based on repeat ground surveys and ASTER satellite imagery. Röhl's (2006) rate of retreat is almost identical (53 m a^{-1}, 2000–2006) to that calculated for the same period during the present study (Tab. 3). This, coupled with the 37% (34 m a^{-1}) lower retreat rate of Warren and Kirkbride (2003), indicates that the mechanisms of ice loss occurring at the terminus of Tasman Glacier have remained reasonably consistent over the period 1994–2006. However, this does not necessarily imply that the mechanisms of ice loss have been simplistic. On the contrary, due to the relatively slow velocities recorded at the glacier terminus over this period, the identification of controlling factors is all the more difficult as the limnological characteristics of Tasman Lake (Kirkbride and Warren 1997; Röhl 2006) and ice-wall melt (Kirkbride and Warren 1999; Warren and Kirkbride 2003) have begun to interact (Röhl 2006).

For the period 2000–2003, Röhl (2005) indicated that processes resulting in the retreat of the Tasman Glacier terminus occurred due to several forms of small-scale calving processes, such as those described by Kirkbride and Warren (1997) at the terminus of Maud Glacier. Broadly these calving processes are classified as flake-calving, small-and large-scale waterline calving, and slab-calving (subaerial). In addition, subaqueous calving was identified as an important ablative process.

The mechanism controlling subaerial calving proposed by Röhl (2005) during the period of retreat 2000 to 2003 is thermal undercutting. Via this process, a thermo-erosional notch develops at the base of the ice cliff at the waterline, destabilising the subaerial cliff causing sections to calve off (Röhl 2006). Thermal undercutting during this period occurred at a rate of between 0.1 to 0.3 m d^{-1} during summer, accounting for the retreat of the glacier terminus (Röhl 2006). As the average annual u_r of the glacier terminus did not change between 2003 and 2006, it can be assumed that this process of thermal undercutting and notch development remained the primary cause of glacier retreat during this period.

Photo 4: Catastrophic break-up of a large part of the glacier terminus as seen in March 2007 (Photo: S. Winkler)

Tab. 3: Frontal retreat and retreat rate, based on sequential satellite images, for the Tasman Glacier terminus over the period 2000–2008. Δx is frontal retreat, u_r is retreat rate ($\Delta x/\Delta t$), u_r width and annually averaged. The values in the rows are for the period starting with the date on the previous line ending with the date on the current line

Date	Time interval: days (Δt)	Δx_{max} (m)	Δx_{mean} (m)	u_r (m a^{-1})
29/04/2000				
07/04/2001	344	232	66 ± 29	72
14/02/2002	313	117	24 ± 28	38
31/12/2002	321	84	37 ± 18	47
09/09/2005	984	187	88 ± 59	33
29/04/2006	233	501	108 ± 144	145
23/01/2007	269	1651	190 ± 366	258
08/04/2007	75	399	46 ± 76	250
13/05/2008	401	297	73 ± 77	123
18/10/2008	159	273	48 ± 66	169
2000-2006	2195		323	54
2006-2008	904		357	144

Large-scale changes occurred at the terminus of Tasman Glacier resulting in rapid retreat between 2006 and 2008. u_r for the glacier increased significantly compared to the previous period of retreat, with full width mean retreat increasing from 54 m a^{-1} for 2000–2006 to 144 m a^{-1} during 2006–2008. This increase in retreat rate between the periods signifies a transition between the principal controlling factors acting at the terminus of the glacier. The probable reason for this is the retreat of the calving ice cliff into deeper water as the glacier retreats up-valley into a deeper part of the U-shaped glacial valley. Indeed, a global data set compiled by Haresign (2004), shows that calving rate is a function of increasing water depth, with rates an order of magnitude higher for tidewater glaciers (Benn et al. 2007).

A secondary factor affecting terminus dynamics is the development and growth of several supraglacial ponds as a result of the slow moving, low gradient nature of the lower glacier tongue. The expansion of supraglacial ponds occurred throughout 2000–2008, however, their impact on terminus dynamics only became evident during 2006–2008 when expanded to c. 583,000 m^2 (including ponds connected in planform to Tasman Lake). This indicates a significant increase in pond surface area, immediately prior to the period of rapid retreat.

The growth of supraglacial ponds over this period significantly contributed to the alteration of the stress regime at the glacier terminus, by increasing the buoyancy of this portion of the glacier (Röhl, 2008). This is due to the significant difference in mass between water-filled and ice-filled ponds, owing to the proportion of air filling the depression between the pond level and the top of the surrounding ponds ice cliff. For example, if the volume of a depression occupied by a pond and air were substituted with ice, the overburden weight would increase in the order of 10 t m^{-2} (Röhl 2008). This is important because as a grounded water-terminating glacier terminus becomes buoyant by the expansion of supraglacial ponds, the rate of ice loss increases substantially as a result of changes in glacier dynamics and stress regimes (Warren et al. 2001; van der Veen 2002; Röhl 2008).

The expansion of supraglacial ponds through to December 2006 increased the buoyancy of Tasman Glacier to the point where a threshold was met and the terminus disintegrated in early 2007 (Photos 3, 4). This is probably due to similar processes of buoyancy-driven calving as described by Warren et al. (2001) for Glacier Nef, Chile, where the calving of large 'tabular' icebergs as a result of hinge-calving mechanisms occurred. Buoyant forces are thought to impose torque on the glacier terminus, causing the release of coherent sections of the glacier front due to crack-propagation at the glacier bed. Such large coherent icebergs are clearly identifiable in photos 3 and 4 during the disintegration of the lower glacier tongue, providing evidence of this process as a likely scenario.

At the moment, Tasman Lake is by far the largest proglacial/supraglacial lake in the Southern Alps. Whereas the rate of terminal retreat and lake en-

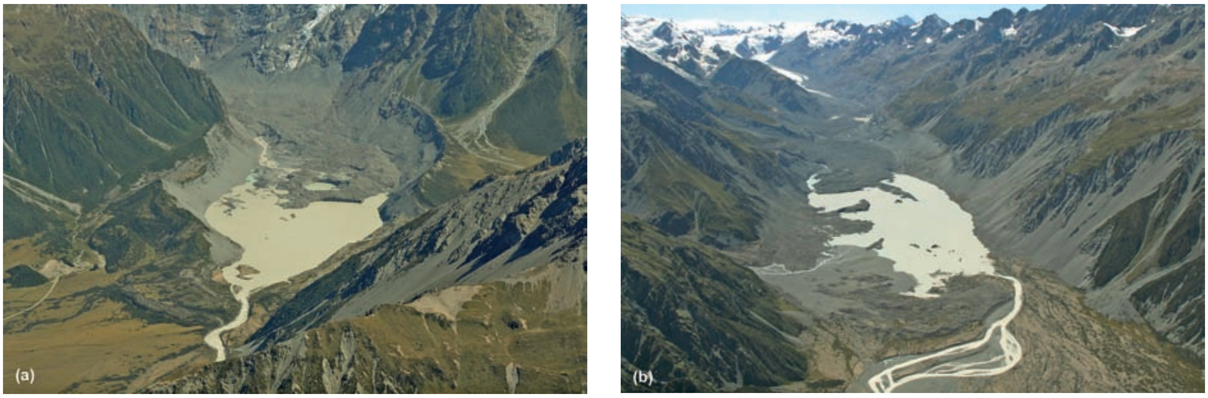

Photo 5: Oblique air photos of Mueller Glacier (a) and Murchison Glacier (b), both now entering a period of rapid lake enlargement (Photos: S. Winkler, March 2008)

largement slowed down at some of the glaciers (e.g. Godley, Hooker and Ramsay Glaciers) reaching a form of balance and/or responding to positive net balances during the past few decades (Chinn et al. 2005, 2008), the development at Tasman Lake is, as shown here, different. Mueller and Murchison glaciers, on the other hand, have just entered a period of fast lake growth and will most certainly experience comparable fast retreat rates in the near future (Photo 5). This development is an impressive example of how debris-covered glaciers with proglacial lakes have uniquely dynamic processes partly or completely decoupled from the climate and mass balance trend of a region (cf. Winkler et al. this issue).

6 Conclusion

This study has documented the retreat of Tasman Glacier and the evolution of Tasman Lake, utilising multispectral satellite imagery for the period 2000–2008 and a bathymetric survey of Tasman Lake undertaken in early April 2008. From the quantification of observed retreat of Tasman Glacier, and Tasman Lake bathymetry, the identification of the key processes operating at the terminus of Tasman Glacier and within Tasman Lake has been undertaken.

The retreat of Tasman Glacier has been rapid, although significant temporal variations in the rate of retreat have also been identified. Specifically, there is an identifiable transition in the rate of retreat prior to, and after, 2006. Retreat prior to 2006 represented a continuation of slow retreat primarily as a result of low magnitude-high frequency iceberg calving, similar to that identified in previous studies conducted at Tasman Glacier. Iceberg calving during this period was for the most part controlled by thermal undercutting at the waterline of the Tasman Glacier ice cliff. After 2006 retreat accelerated significantly as terminus dynamics were altered. Although it is difficult to identify the controlling mechanisms from a retrospective study, available evidence points to buoyancy-driven calving as the primary calving mechanism for the lower terminus.

The result of Tasman Glacier retreat has been the expansion of Tasman Lake. Significant increases in lake surface area and volume have occurred as Tasman Lake has expanded into deeper water. The highly irregular bathymetry near the terminus of Tasman Glacier indicates an ice-cored lake floor.

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References

Anderson, B.; Lawson, W.; Owens, I. and Goodsell, B. (2006): Past and future mass balance of 'Ka Roimata o Hine Hukatere' Franz Josef Glacier, New Zealand. In: Journal of Glaciology 52, 597–607. DOI: 10.3189/172756506781828449

Anderton, P. W. (1975): Tasman Glacier 1971–73. Hydrological Research Annual Report 33, Wellington.

Bamber, J. L. and Payne, A. J. (eds.) (2004): Mass balance of the cryosphere. Observations and modelling contemporary and future changes. Cambridge.

Benn, D. I.; Warren, C. R. and Mottram, R. H. (2007): Calving processes and the dynamics of calving glaciers. In: Earth Science Reviews 82, 143–179. DOI: 10.1016/j.earscirev.2007.02.002

Blair, R. W. (1994): Moraine and valley wall collapse due to rapid deglaciation in Mount Cook National Park, New Zealand. In: Mountain Research and Development 14, 347–358. DOI: 10.2307/3673731

Broadbent, M. (1973): A preliminary report on seismic and gravity surveys on the Tasman Glacier 1971–2. Wellington.

Brook, M. S.; Kirkbride, M. P. and Brock, B. W. (2006): Quantified time scale for glacial valley cross-profile evolution in alpine mountains. In: Geology 34, 637–640. DOI: 10.1130/G22700.1

Burrows, C. J. (1973): Studies on some glacial moraines in New Zealand. 2. Ages of moraines of the Mueller, Hooker and Tasman Glaciers. In: New Zealand Journal of Geology and Geophysics 16, 831–855.

– (2005): Julius Haast in the Southern Alps. Christchurch.

Chinn, T. (1996): New Zealand glacier responses to climate change of the past century. In: New Zealand Journal of Geology and Geophysics 39, 415–428.

Chinn, T.; Winkler, S.; Salinger, M. J. and Haakensen, N. (2005): Recent glacier advances in Norway and New Zealand: A comparison of their glaciological and meteorological causes. In: Geografiska Annaler 87 A, 141–157.

Chinn, T.; Salinger, J.; Fitzharris, B. B. and Willsman, A. (2008): Glaciers and climate. In: Bulletin of the Federal Mountain Clubs of NZ 171, 1–15.

Claridge, D. (1983): A geophysical study of the termini of the Mount Cook National Park glaciers. Unpubl. MSc thesis, University of Auckland, Auckland.

Drewry, D. (1986): Glacial geologic processes. London.

Gellatly, A. F. (1982): Lichenometry as a relative-age dating method in Mount Cook National Park, New Zealand. In: New Zealand Journal of Botany 20, 343–353.

– (1985): Historical records of glacier fluctuations in Mt Cook National Park, New Zealand: a century of change. In: The Geographical Journal 151, 86–99. DOI: 10.2307/633283

Gustavson, T. C. (1975): Bathymetry and sediment distribution in proglacial Malaspina Lake, Alaska. In: Journal of Sedimentary Petrology 45, 450–461.

Hambrey, M. J. and Ehrmann, W. (2004): Modification of sediment characteristics during glacial transport in highalpine catchments: Mount Cook area, New Zealand. In: Boreas 33, 300–318. DOI: 10.1080/03009480410001965

Haresign, E. C. (2004). Glacio-limnological Interactions at Lake-calving Glaciers. Unpubl. PhD Thesis, University of St Andrews, UK.

Hoelzle, M.; Chinn, T.; Stumm, D.; Paul, F.; Zemp, M. and Haeberli, W. (2007): The application of glacier inventory data for estimating past climate change effects on mountain glaciers: a comparison between the European Alps and the Southern Alps of New Zealand. In: Global and Planetary Change 56, 69–82. DOI: 10.1016/j.gloplacha.2006.07.001

Hochstein, M. P., Claridge, D., Henrys, S. A., Pyne, A., Nobes, D. C. and Leary, S. F. (1995): Down-wasting of the Tasman Glacier, South Island, New Zealand: changes in the terminus region between 1971 and 1993. In: New Zealand Journal of Geology and Geophysics 38, 1–16.

IPCC (ed.) (2007): Climate change 2007: the physical science basis. Contribution of working group I to the 4th Assessment report of the Intergovernmental Panel on Climate Change. Cambridge.

Jensen, J. R. (2007): Remote sensing of the environment: an earth resource perspective. New Jersey.

Kääb, A. (2002): Monitoring high-mountain terrain deformation from repeated air- and spaceborne optical data: examples using digital aerial imagery and ASTER data. In: Journal of Photogrammetry and Remote Sensing 57, 39–52. DOI: 10.1016/S0924-2716(02)00114-4

Kirkbride, M. P. (1989): The influence of sediment budget on geomorphic activity of the Tasman Glacier, Mount Cook National Park, New Zealand. Unpubl. PhD thesis, University of Canterbury, Christchurch.

– (1993): The temporal significance of transitions from melting to calving termini at glaciers in the Southern Alps of New Zealand. In: The Holocene 3, 232–240. DOI: 10.1177/095968369300300305

– (2000): Ice marginal geomorphology and Holocene expansion of debris-covered Tasman Glacier, New Zealand. In: Nakawo, M.; Raymond, C.F. and Fountain, A. (eds.): Debris-Covered Glaciers. IASH Publication 264, 211–218.

Kirkbride, M. P. and Brazier, V. (1998): A critical evaluation of the use of glacier chronologies in climatic reconstruction, with reference to New Zealand. In: Quaternary Proceedings 6, 55–64.

Kirkbride, M. P. and Spedding, N. (1996): The influence of englacial drainage on sediment transport pathways and till texture of temperate valley glaciers. In: Annals of Glaciology 22, 160–166.

Kirkbride, M. P. and Warren, C. R. (1997): Calving processes at a grounded ice cliff. In: Annals of Glaciology 24, 116–121.

– (1999): Tasman Glacier, New Zealand: 20th-century thinning and predicted calving retreat. In: Global and Planetary Change 22, 11–28. DOI: 10.1016/S0921-8181(99)00021-1

Knight, P. G. (ed.) (2006): Glacier science and environmental change. Oxford.

Lendenfeld, R. v. (1884): Der Tasman-Gletscher und seine Umrandung. In: Petermanns Mitteilungen Ergänzungsband XVI, 1–80.

Nakawo, M.; Raymond, C. F. and Fountain, A. (eds.) (2000): Debris-covered glaciers. IAHS Publication 264. Wallingford.

Purdie, J. and Fitzharris, B. B. (1999): Processes and rates of ice loss at the terminus of Tasman Glacier, New Zealand. In: Global and Planetary Change 22, 79–91. DOI: 10.1016/S0921-8181(99)00027-2

Purdie, H. L.; Brook, M. S. and Fuller, I. C. (2008): Seasonal variation in ablation and velocity on a temperate maritime glacier: Fox Glacier, New Zealand. In: Arctic, Antarctic, and Alpine Research 40, 140–147. DOI: 10.1657/1523-0430(06-032)[PURDIE]2.0.CO;2

Reynolds, J. M. (2000): On the formation of supraglacial lakes on debris-covered glaciers. In: Nakawo, M.; Raymond, C. F. and Fountain, A. (eds.): Debris-covered glaciers. IASH Publication 264, 211–218.

Röhl, K. (2005): Terminus disintegration of debris-covered, lake-calving glaciers. Unpubl. PhD thesis, University of Otago, Dunedin.

– (2006): Thermo-erosional notch development at fresh-water-calving Tasman Glacier, New Zealand. In: Journal of Glaciology 52, 203–212. DOI: 10.3189/172756506781828773

– (2008): Characteristics and evolution of supraglacial ponds on debris-covered Tasman Glacier, New Zealand. In: Journal of Glaciology 54, 867–880. DOI: 10.3189/002214308787779861

Röthlisberger, F. (1986): 10.000 Jahre Gletschergeschichte der Erde. Aargau.

Schaefer, J. M.; Denton, G. H.; Kaplan, M.; Putnam, A.; Finkel, R. C.; Barrell, D. J. A.; Andersen, B. G.; Schwartz, R.; McIntosh, A.; Chinn, T. and Schlüchter, C. (2009): High-frequency Holocene glacier fluctuations in New Zealand differ from the northern signature. In: Science 324, 622–625. DOI: 10.1126/science.1169312

Sugden, D. E. and Kirkbride, M. P. (1992): New Zealand loses its top. In: Geographical Magazine 64, 30–34.

Van der Veen, C. J. (2002): Calving glaciers. In: Progress in Physical Geography 26, 96–122. DOI: 10.1191/0309133302pp327ra

Von Haast, I. (1866): The glaciers on Mount Cook and Mount Tasman (sketchmap 1 inch: 4 miles). Wellington.

Warren, C. R. and Kirkbride, M. P. (2003): Calving speed and climate sensitivity of New Zealand lake-calving glaciers. In: Annals of Glaciology 36, 173–178. DOI: 10.3189/172756403781816446

Warren, C. R.; Benn, D.; Winchester, V. and Harrison, S. (2001): Buoyancy-driven lacustrine calving, Glaciar Nef, Chilean Patagonia. In: Journal of Glaciology 47, 135–146. DOI: 10.3189/172756501781832403

Watson, M. I. (1995): Geophysical and glaciological studies of the Tasman and Mueller Glaciers. Unpubl. MSc thesis, University of Auckland, Auckland.

Winkler, S. (2004): Lichenometric dating of the 'Little Ice Age' maximum in Mt Cook National Park, Southern Alps, New Zealand. In: The Holocene 14, 911–920. DOI: 10.1191/0959683604hl767rp

– (2005): The Schmidt hammer as a relative-age dating technique: potential and limitations of its application on Holocene moraines in Mt Cook National Park, Southern Alps, New Zealand. In: New Zealand Journal of Geology and Geophysics 48, 105–116.

– (2009): Gletscher und ihre Landschaften – eine illustrierte Einführung. Darmstadt.

Winkler, S.; Chinn, T.; Gärtner-Roer, I.; Nussbaumer, S. U.; Zemp, M. and Zumbühl, H. J. (2010): An introduction to mountain glaciers as climate indicators with spatial and temporal diversity. In: Erdkunde 64 (2), 97–118. DOI: 10.3112/erdkunde.2010.02.01

Zemp, M.; Roer, I.; Kääb, A.; Hoelzle, M.; Paul, F. and Haeberli, W. (2008): Global glaciers changes: facts and figures. Zürich.

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