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Source: Arctic, Antarctic, and Alpine Research, 43(1) : 1-10

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1938-4246-43.1.1

Arctic, Antarctic, and Alpine Research, Vol. 43, No. 1, 2011, pp. 1–10

Twenty-First Century Calving Retreat of Tasman Glacier, Southern Alps, New Zealand

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Abstract

Tasman Glacier is the largest glacier in the New Zealand Southern Alps. Despite a century of warming and down-wastage, the glacier remained at its Little Ice Age terminus until the late 20th century. Since then, a proglacial lake formed, and comparatively rapid calving retreat has been initiated. In this paper we use sequential satellite imagery to document terminus retreat, growth of supraglacial ponds, and expansion of the proglacial Tasman Lake. Between 2000 and 2008, the glacier terminus receded a maximum of c. 3.7 km on the western margin, and the ice-contact Tasman Lake expanded concomitantly. This northward expansion of Tasman Lake up-valley proceeded at a mean annual rate of 0.34×10^6 m^2 a^{-1} over 2000–2008, attaining a surface area of 5.96×10^6 m^2 in May 2008, with a maximum depth of c. 240 m. Terminus retreat rates (U_r) vary in both space and time, with two distinct periods of calving retreat identified during the study period: 2000–2006 (mean U_r = 54 m a^{-1}) and 2007–2008 (mean U_r = 144 m a^{-1}). Terminus retreat can also be categorized into two distinct zones of activity: (1) the main ice cliff (MIC), and (2) the eastern embayment ice cliff (EEIC). During the period 2000–2006, and between 2006 and 2008 for the EEIC, the controlling process of ice loss at the terminus was iceberg calving resulting from thermal undercutting. In contrast, the retreat of the MIC between 2006 and 2008 was controlled by buoyancy-driven iceberg calving caused by decreasing overburden pressure as a result of supraglacial pond growth, increased water depth, and rainfall. The presence of a >130-m-long subaqueous ice ramp projecting from the terminal ice cliff into the lake suggests complex interactions between the glacier and ice-contact lake during the 8–10 km of possible future calving retreat.

DOI: 10.1657/1938-4246-43.1.1

Introduction

Recession and down-wasting of glaciers in alpine mountains and formation of water-terminating glacier margins is receiving an increasing amount of attention in the literature, as glaciers respond to climate warming and terminus retreat prevails (Benn et al., 2007). Such 'calving' glacier margins have evolved in regions such as Patagonia (Warren et al., 2001; Skvarca et al., 2002), the Himalaya (Benn et al., 2001), the European Alps (Huggel et al., 2002; Diolaiuti et al., 2006), and New Zealand (Warren and Kirkbride, 2003; Röhl, 2008). Calving is generally assumed to be the mechanical loss of ice from glaciers and ice shelves, though it has become increasingly apparent that melting, both subaerial and subaqueous, is a significant process at calving termini, about which little is known. Calving is an important part of the mass budget of many glacier systems globally, and is a particularly important process to glaciologists because the dynamic behavior of calving glaciers may be partially decoupled from climate, with factors other than variation in the equilibrium line altitude being a major control on calving terminus location and rate of advance/retreat. One particular variable that appears important is water depth, with several authors identifying a positive relationship between calving rate and increasing water depth (e.g., Pelto and Warren, 1991; Warren et al., 1995; Warren and Kirkbride, 2003). Nevertheless, calving glaciers can undergo very rapid retreat following a climate signal, and so an understanding of ablative mechanisms such as calving is important for predicting glacier response to climatic forcing. As with many glaciological problems, examining calving mechanisms, and the processes that control variation in calving rates, is compounded by the difficulties and dangers of the research environment.

One process that has remained elusive to such analysis is subaqueous calving (Warren and Kirkbride, 1998), and specifically the presence or absence of a submerged projecting ice "foot" or "ramp," extending from the calving ice cliff. Early observations of icebergs rising to the water surface tens to hundreds of meters in front of Alaskan tidewater glaciers led Russell (1891) and Wright (1892) to postulate a projecting ramp of ice extending as much as 300 m beyond the subaerial ice cliff, an idea later supported by Tarr (1909). Reid (1892), however, argued that the submerged parts of the termini of these glaciers were vertical or even undercut, and suggested that blocks of ice which broke off travelled laterally underwater before breaking the surface. Since these early reports, only a paucity of data has been collected on ice ramps due to the inherent dangers of investigating this problem, and this has prevented resolution of the debate. Nevertheless, discussion has recently been revived given the general interest in retreat of calving glaciers (e.g., Motyka, 1997), because resolving the question of calving geometry would enhance understanding of calving controls, the significance of subaqueous melt, and the relationships between calving and glacier dynamics (Röhl, 2006). For example, it has recently been suggested (O'Neel et al., 2007) that the close

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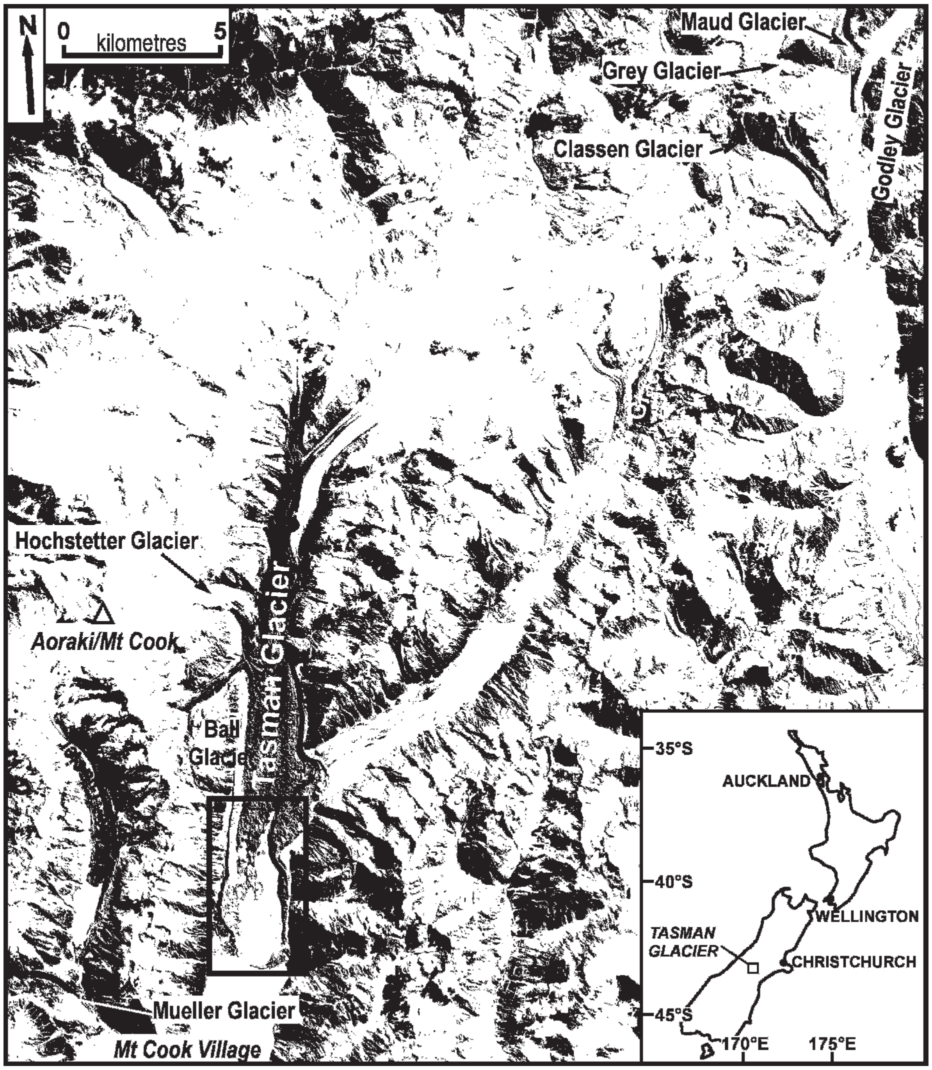

FIGURE 1. Location of Tasman Glacier in the *Aoraki*/Mt Cook National Park, and location of other ice-contact lakes and glaciers mentioned in the text. The base map is an ASTER image acquired on 9 February 2006. Box around Tasman Lake represents the area shown in Figures 2 and 3A.

association between subaerial and subaqueous calving events reflects the sudden reduction of ice overburden pressure on the submerged portion, and the consequent increase in buoyant forces.

In this paper we use a range of remotely sensed data to examine recent (2000–2008) retreat of Tasman Glacier, New Zealand. This is a debris-covered lake-calving glacier that has undergone dramatic changes over the last two decades. The aims are threefold: (1) to document the 21st century retreat of Tasman Glacier through the use of multispectral satellite imagery of calving terminus position; (2) to map lake bathymetry via echo-sounding; and (3) to determine the presence/absence of subaqueous ice ramps projecting from the calving terminus using towfish sonar sub-surface profiling.

Study Area

Tasman Glacier (43°35′S, 170°10′E), with a length of ~29 km, is New Zealand's longest glacier (Fig. 1), draining the eastern side of the Southern Alps in *Aoraki*/Mount Cook National Park. Debris cover extends from the terminus to ~8 km up-glacier, with thickness increasing down-glacier to ~1 m at the terminus (Röhl, 2008), but it can be variable. Initial ice loss over the 20th century was by down-wasting, with Skinner (1964) estimating a vertical loss of 82 m of ice near the Ball Glacier confluence between 1890 and 1962. More recently, Hochstein et al. (1995) have indicated a down-wasting rate of between 0.3 and 1.2 m a^{-1}. Decades of down-wasting since the mid-20th century led to the development of numerous supraglacial ponds, and during the 1980s a proglacial lake (at 727 m a.s.l.) began to form by coalescence of ponds (Kirkbride and Warren, 1999). The transition to a calving terminus followed in the 1990s, and in 2003 the lake had a maximum depth of 185 m, covering an area of 3.7×10^6 m^2 (Röhl, 2008). Ice loss at the terminus is now dominated by calving into the proglacial lake (Purdie and Fitzharris, 1999), with the glacier currently retreating from its maximum position, attained during the Little Ice Age (Chinn, 1996). Recent work by Quincey and Glasser (2009) has used remotely sensed data to derive glacier velocities, rates of surface lowering, and lake expansion. The general patterns they report are lake growth proceeding at increasing rates—for example, the lake surface area doubling between 2000 and 2007. Rather than being impounded by terminal moraine, the lake basin is developing behind a large Holocene outwash head (Quincey and Glasser, 2009), removing the possibility of catastrophic moraine dam failure. Considerable work has been undertaken on the terminus dynamics and lake properties (e.g., Hochstein et al., 1995; Kirkbride and Warren, 1999; Purdie and Fitzharris, 1999; Röhl, 2006, 2008), with complex interactions between limnological and glaciological processes apparent. The freeboard of the subaerial ice cliff varies between 10 and 15 m high (Röhl, 2008), though little is known about the subaqueous geometry of the glacier terminus. From studying thermo-erosional notch development at the water line of the calving ice cliff, Röhl (2006) found that calving rates are directly controlled by the rate of thermal undercutting. Warren and Kirkbride (1998) favored the idea of a planar subaqueous ramp at Tasman Glacier, suggesting a minor ramp <4 m wide might be formed from debris-mantled extensions of ice. In contrast, Hochstein et al. (1995) favored the idea of a vertical calving cliff in most parts, descending >200 m to the lake bed. In a slight contrast, Purdie and Fitzharris (1999) suggested that narrow, low gradient ramps may form at Tasman Glacier due to rapid subaerial calving, but that any subaqueous ice ramp would melt back 'fairly rapidly' to match the freeboard profile (Purdie and Fitzharris, 1999, p. 87).

Regional climate in the Southern Alps is dominated by the prevailing southwesterly airflow. Precipitation is orographically enhanced by the Southern Alps, resulting in highest annual rainfalls of 15,000 mm a^{-1} west of the Main Divide (Chinn et al., 2005). Nevertheless, precipitation is high in the Tasman Glacier area, varying from approximately 7000 mm a^{-1} in the higher elevations (Anderton, 1975) to around 3000 to 5000 mm a^{-1} near the terminus (Röhl, 2006). Examination of linkages between climate and glacier change using atmospheric circulation (Fitzharris et al., 1997; Chinn et al., 2005) shows that regional circulation patterns exert a strong control on glacier balance across the Southern Alps. Indeed, mass changes indicate strong correlations with the El Niño–Southern Oscillation (ENSO) and the 18–20 year variability of the Inter-decadal Pacific Oscillation (Chinn et al., 2005). Positive mass balances and advances are associated with an increase in the prevailing west-southwest airflow. Negative mass balances and retreats are associated with more northeasterly flow (Chinn et al., 2005). Such retreats and advances are especially visible on the short response time (6–9 year) of Fox and Franz Josef glaciers to the west of the Main Divide, which advanced during the 1980s and 1990s (Purdie et al., 2008) due to negative ENSO-driven precipitation increases of 10–15% (Chinn et al., 2005). These ~11-km-long glaciers are steep, lack any insulating debris cover, and terminate <300 m a.s.l. In contrast, glaciers such as Tasman Glacier are much longer (>25 km), lower gradient, and are debris-mantled, and so have a more delayed terminus response to mass balance change, of the order of a century (Chinn et al., 2005).

Methods

CALVING TERMINUS POSITION

In order to delineate the Tasman Glacier calving terminus position and Tasman Lake area, multispectral satellite images were collected for the period of 29 April 2000 to 18 October 2008 from ASTER, SPOT, and Quickbird. Satellite images were georeferenced using an 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 (1949), was used as the base image from which all other images were registered. All other images were registered using the rotation, sampling and translation (RST) warping method (e.g. Jensen, 2007), fixed to over 20 GCPs. The resulting horizontal root mean square error (RMS) for each rectified image was smaller than a single pixel. The resulting accuracy of the digitized terminus positions from the satellite images is estimated at ~15 m (Röhl, 2008).

Terminal ice cliff retreat rates (U_r) in this study were calculated by measuring the perpendicular distance between terminus positions (Fig. 2), evenly spaced at 100 m intervals along the ice front. The change in distance along the terminus (∂L) was then annually averaged to provide a rate of change (U_r) for each measured location during the study period, which was then width-averaged. From the digitized 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.

LAKE BATHYMETRY AND SUBAQUEOUS GEOMETRY

Lake bathymetry was mapped in April 2008 using a Humminbird 323 DualBeam Plus dual frequency echo-sounder (operating frequencies of 200 kHz and 83 kHz) and a hand-held GPS unit. Accuracy of water-depth (z) measurements is typically ±6 m, and point position (x, y) accuracy for the GPS unit is known to be ±5 m. 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 calving ice cliff, due to subaqueous geometry, so echo-sounder readings were limited to distances >50 m from the ice cliff. Bathymetric data was then interpolated using triangular irregular network (TIN) modeling from which smoothed bathymetric contours were produced (Fig. 2). To determine whether an ice ramp is present at Tasman Glacier, two seismic profiles were obtained, one parallel to the subaerial ice cliff and one normal to the ice cliff, oriented north-south into the proglacial lake (locations on Fig. 2). Traces were collected by towing an EdgeTech 216S towfish sub-bottom profiler, via a 25-m-long cable, attached to a 3 m boat. The frequency range of the EdgeTech 216S is 2–16 kHz, giving a vertical resolution of 6–10 cm (EdgeTech, 2005).

Results

TERMINUS RETREAT AND CALVING RATES

Retreat of the terminus position (U_r) of Tasman Glacier has been substantial during 2000–2008 (Figs. 2 and 3), though significant temporal and spatial variations are also evident. Retreat rate was at a minimum during December 2002 to September 2005, with a mean U_r of 33 m a^{-1} (Table 1). U_r peaked during the interval April 2006 to January 2007, at 497 m a^{-1}. In more detail, there appears to be an identifiable transition in the rate of retreat prior to, and after, 2006. Also, terminus retreat of Tasman Glacier can be broadly categorized into two distinct zones of activity, (1) the rapidly retreating main ice cliff (MIC), and (2) the relatively stable eastern embayment ice cliff (EEIC) that extends up the true left of the glacier snout (Figs. 2 and 3B). The rate of retreat for the entire terminal ice cliff, the MIC, and the EEIC are given in Table 1.

Looking at the two periods of activity, pre-2006 and 2006 onwards, during 2000–2006 the terminal ice cliff retreated a mean distance (∂L_{mean}) of 323 m, equating to a retreat rate (U_r) of 54 m a^{-1} (Table 1). Within this period, the actual frontal retreat was highly varied for each survey interval, and, consequently, U_r has ranged between 33 and 226 m a^{-1}, indicating wide temporal variability in terminus retreat. The full-width data for 2000–2006, however, masks a marked spatial variation in retreat rate (U_r). Indeed, retreat data from the MIC and EEIC (Table 1) indicate greater change occurring along the MIC (Figs. 2 and 3B), with retreat rates varying between 37 and 255 m a^{-1} and 8 and 47 m a^{-1} for the MIC and EEIC, respectively.

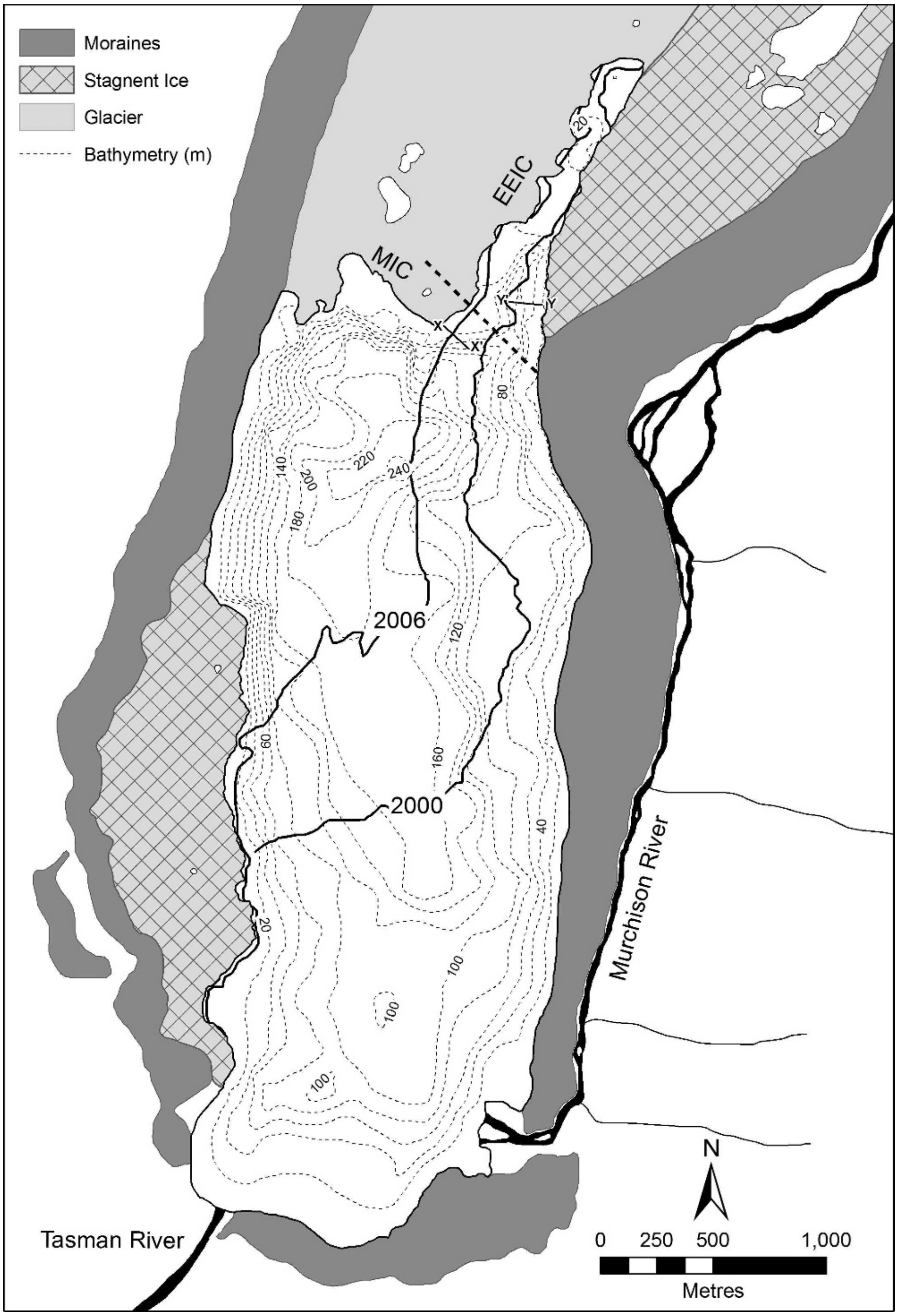

FIGURE 2. Bathymetric map of Tasman Lake from the April 2008 echo-sounding survey. The location of the terminus in 2000 and 2006 is shown along with two sub-bottom sonar profiles (Fig. 6). The subdivision between the eastern embayment ice cliff (EEIC) and the main ice cliff (MIC), north and south of the dotted line, is also shown.

From 2006 onwards, retreat rate increased dramatically, with full-width mean retreat (U_r) of 144 m a^{-1} between 2006 and 2008 (Table 1), significantly higher than the mean 2000–2006 U_r of 54 m a^{-1} (Table 1). As with the pre-2006 study period, from 2006 onwards, U_r has varied between study intervals too, ranging between 66 and 497 m a^{-1} (Table 1). During this period, maximum terminus retreat occurred during a 9 month period between April 2006 and January 2007, when the glacier retreated c. 1600 m on the true right (western) side of the terminus (Fig. 3). The spatial variation between the retreat rate of the eastern embayment and the main ice cliff over this period (2006–2008) continued, with U_r rates of 218 and 27 m a^{-1}, for MIC and EEIC, respectively (Table 1). In particular, MIC U_r ranged markedly between 97 and 592 m^{-1}, but was almost constant at the EEIC, with U_r ranging between 11 and 47 m a^{-1} (Table 1).

Following Benn et al. (2007), calving rate, U_c (m a^{-1}), for Tasman Glacier was obtained by calculating the difference between glacier velocity (U_i) at the terminus, and the rate of change in glacier length (∂L) over a given period of time (∂t):

$$U_c = U_i - \frac{\partial L}{\partial t}. \quad (1)$$

Calving data for the full width, MIC and EEIC of Tasman Glacier is presented in Table 1. Ice velocity (U_i) and calving cliff melt rate (U_m) are utilized from Röhl (2006). The term *calving rate*

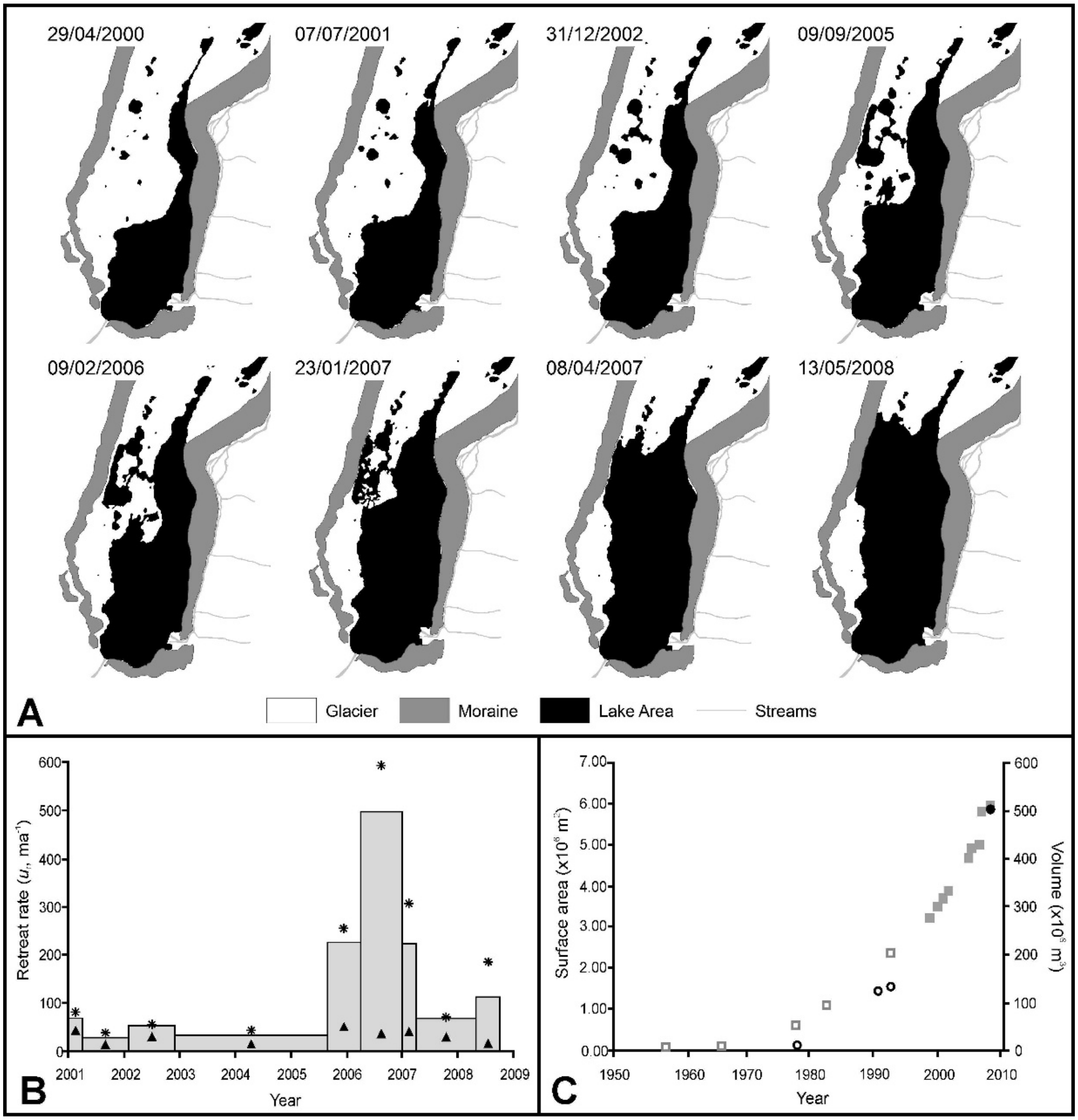

FIGURE 3. (A) Growth of Tasman Lake and supraglacial ponds between April 2000 and May 2008. (B) Graph of Tasman Glacier retreat showing the mean (bars), MIC (asterisks), and EEIC (triangles) retreat rates. (C) Surface area (squares) and water volume (circles) of meltwater ponds (pre-1980s) and Tasman Lake between 1957 and 2008. Solid points are from this study, surface area and lake volume pre-2000 from Kirkbride (1989), Hochstein et al. (1995), and Watson (1995).

hereafter refers to $U_c + U_m$, because the melting of the calving face should be incorporated into calculations (Warren and Kirkbride, 2003). Following this, full-width calving rates (U_c) follow the same temporal variations and spatial contrasts as the retreat rates (U_r) outlined above, with full-width U_c lowest (32 m a^{-1}) between December 2002 and September 2005, and peaking at 501 m a^{-1} during April 2006 to January 2007. Spatial contrasts reveal that from 2000 to 2005, U_c for MIC varied from 46 to 79 m a^{-1}, with a mean U_c of 78 m a^{-1}, similar orders of magnitude U_c values to the EEIC, which varied from 11 to 45 m a^{-1} (mean U_c = 24 m a^{-1}, Table 1). During 2006 to 2008 the mean annual calving rate (U_c) for the MIC was 227 m a^{-1}, with a maximum U_c of 600 m a^{-1}, significantly larger that the mean and maximum values of 30 m a^{-1} and 49 m a^{-1} for EEIC, respectively (Table 1). This contrast reflects a difference in the controlling mechanisms acting at different locations along the terminal ice cliff.

LAKE BATHYMETRY AND ICE RAMP GEOMETRY

Tasman Lake deepens gradually up-valley of the LIA moraines until it increases in depth dramatically c. 800 m from the current glacier terminus (Fig. 2). From this point onwards, the lake shallows rapidly towards the glacier front indicating the presence of a subaqueous ice ramp. 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 center of the glacier terminus (Fig. 2), located within a trough that runs along the center-right of the lake. In general, the lake floor follows the typical pattern of glacial valley cross-profile morphology. Recession of the calving terminus of Tasman Glacier has resulted in concomitant northward expansion of Tasman Lake up-valley, with the surface area of the lake expanding at a mean rate of 0.34×10^6 m^2 a^{-1} over

TABLE 1

Tasman Glacier retreat data, 2000–2008.

Survey interval	Time interval (t) in days	Full width			MIC			EEIC		
		∂L_{mean} (m)	U_r (m a^{-1})	U_c (m a^{-1})	∂L_{mean} (m)	U_r (m a^{-1})	U_c (m a^{-1})	∂L_{mean} (m)	U_r (m a^{-1})	U_c (m a^{-1})
29/04/00–07/04/01	344	66 ± 29	70	75	74 ± 26	79	87	40 ± 24	42	45
07/04/01–14/02/02	313	24 ± 28	28	32	32 ± 25	37	46	7 ± 27	8	11
14/02/02–31/12/02	321	37 ± 18	53	58	42 ± 18	48	56	26 ± 16	30	32
31/12/02–09/09/05	984	88 ± 59	33	37	115 ± 38	43	51	23 ± 52	9	11
09/09/05–29/04/06	233	108 ± 144	226	230	149 ± 163	255	264	30 ± 13	47	49
29/04/06–23/01/07	269	190 ± 366	497	501	291 ± 436	592	600	25 ± 13	34	36
23/01/07–08/04/07	75	46 ± 76	224	228	63 ± 86	307	315	8 ± 20	39	41
08/04/07–13/05/08	401	73 ± 77	66	71	107 ± 88	97	106	29 ± 14	26	29
13/05/08–18/10/08	159	48 ± 66	110	115	80 ± 63	184	192	5 ± 36	11	14
2000–2006	2195	323	54	59	412	69	78	126	21	24
2006–2008	904	357	144	149	541	218	227	67	27	30

Notes: MIC is main ice cliff, EEIC is eastern embayment ice cliff, ∂L_{mean} is mean retreat, U_c is calving rate, and U_r is retreat rate multiplied-up by t to determine annual rate.

2000–2008 to a total area of 5.96 0.34 $\times$ 10^6 m^2 (May 2008). Lake volume has increased markedly as Tasman Glacier has retreated, resulting in the lake attaining a volume of 510 $\times$10^6 m^3 in 2008, an increase of 377 $\times$ 10^6 m^3 between 1995 and 2008 (Fig. 3C). The increase in lake volume is a function of the increase in surface area of Tasman Lake, but also because Tasman Glacier is retreating into a significantly deeper part of the glacially excavated trough (Anderton, 1975). However, a comparison between bathymetric maps produced by Hochstein et al. (1995) and this study (Fig. 2) indicates that water depth has actually decreased by approximately 20 m within the southern end of the lake. Sedimentation within the southern end of Tasman Lake has therefore affected lake volume, but to a lesser extent than the retreat of Tasman Glacier.

Figure 5A is a c. 120-m-long seismic trace collected normal to the terminal ice cliff, from the terminal ice cliff (left end of trace) out into the lake (right end of trace). Several features of interest can be identified. At the western end of the trace (X), the margins of part of an ice ramp that steepens with distance away from the glacier terminus, southeastwards into Tasman Lake, can be seen. This resembles a subaqueous ice 'ramp,' but rather than being a planar, low-gradient ice ramp, the subaqueous ice margin appears to increase in gradient with distance from the terminus. The second feature of interest is the change in tonal intensity at c. 60 m depth, marking the glacier/substrate boundary (left). Right (ice distal) of this point, the lake bed is clearly evident. Figure 5B shows the c. 130-m-long seismic trace from the bounding lateral moraine wall on the eastern margin of the lake, moving westwards into Tasman Lake. Above the strong dark reflector, a strong upper bounding reflector is interpreted as an ice ramp with an undulating surface. The ramp appears to be within c. 50 m of the lake surface even though the trace is several hundred meters from the calving terminus. About halfway along the trace, a ~30-m-high ice ridge rises from the ice ramp surface indicating the ramp has an irregular surface. Together, both traces strongly suggest that rather than being a low gradient, planar feature, the ramp has an irregular, spatially complex surface.

Discussion

SPATIAL AND TEMPORAL VARIABILITY OF CALVING RETREAT

Tasman Glacier has been steadily retreating since the formation of Tasman Lake in the mid-1980s. As identified from the analysis of sequential satellite imagery, two distinct periods of retreat have occurred between 2000 and 2008. The two periods of retreat relate to an initial period of relatively slow retreat between 2000 and 2006, followed by a secondary period of comparatively rapid retreat between 2006 and 2008. However, due to the spatial variation in the rate of frontal retreat (U_r) between the faster retreating main ice cliff (MIC) and the slower eastern embayment ice cliff (EEIC), additional compounding factors and mechanisms must be causing the increased frontal retreat of the MIC. The temporal variability in retreat rates calculated here mirrors the time series of lake area results reported by Quincey and Glasser (2009). Indeed, from an analysis of remotely sensed data between 1990 and 2007, Quincey and Glasser (2009) found that the growth of Tasman Lake is now proceeding at increasing rates post-2006.

Ice velocity contrasts between the MIC and EEIC could explain the difference in retreat rate (U_r) between 2000 and 2006. Ice velocities measured by Röhl (2006) for the MIC and EEIC were 2.5 and 8.5 m a^{-1}, respectively. The lower velocity of the MIC, would allow thermal notching to proceed more vigorously than at the less stationary EEIC. Enhanced thermo-erosional notching would destabilize the ice cliff, leading to larger-scale subaerial calving, increasing the retreat rate (Kirkbride and Warren, 1997; Röhl, 2006).

SUPRAGLACIAL POND GROWTH

A further factor affecting terminus dynamics is the growth of supraglacial ponds since 2000 as the slow moving, low-gradient glacier tongue has continued to down-waste (Fig. 3A). Supraglacial ponds on the debris-covered tongue at the same elevation as Tasman Lake expanded to a combined surface area of c. 583,000 m^2 (including ponds partially connected in planform to Tasman Lake) by 2006. Such supraglacial ponds on the glacier tongue are an important component of terminus destabilization and calving retreat (Röhl, 2008). Ponds that are hydraulically connected (such as those found on the lower terminus of Tasman Glacier) to the main englacial drainage system expand as a result of increased water circulation, contributing to longer term ice loss. Such hydraulically connected ponds grow both laterally and vertically as a result of subaerial and subaqueous melt (Sakai et al., 2000; Benn et al., 2001), increasing pond volume. A limnological threshold specific to each pond is met, controlled by pond temperature and currents, resulting in the initiation of small-scale subaerial calving, similar to that found at the terminus of the glacier (Röhl, 2008). Subaqueous calving is also initiated at a similar threshold, deepening the pond. The result of pond evolution on the lower terminus is an increase in ice loss at pond edges as a result of subaerial calving (Benn et al., 2001). The corollary of this pond expansion is that once ponds begin to drain, the buoyancy of the terminus is increased, altering terminus dynamics and stress regimes (Warren et al., 2001; Van der Veen, 2002; Röhl, 2008). Hence, expansion of supraglacial ponds until 2006 appears to be linked to the rapid terminus retreat post-2006.

TERMINUS BUOYANCY

During 2000–2006, the glacier had receded into significantly deeper waters (Fig. 2), attaining a mean water-depth (h_w) of 153 m (approximated from water-depth measurements along the 2006 ice cliff location [see Fig. 2]). During retreat, the subaerial ice cliff height remained at a relatively constant height of between 10–15 m (Röhl, 2006) as water depth increased to greater than c. 200 m, increasing the ratio between subaerial ice cliff height ('freeboard') and water depth. Torque arising from buoyancy forces acting on the terminus increases as this ratio rises, until a threshold is met where the glacier can no longer maintain a grounded terminus (Vieli et al., 2001; Van der Veen, 2002), as basal shear stress decreases to zero (Warren et al., 2001). The point at which this threshold is met can be defined by the flotation thickness of a glacier, h_f, calculated as:

$$h_f = \frac{\rho_w}{\rho_i} h_w \qquad (2)$$

where ρ_w is water density (1000 kg m^{-3}), ρ_i is the density of glacier ice (917 kg m^{-3}), and h_w is water depth. For the MIC prior to the disintegration of the terminus h_w was c. 153 m, giving a h_f of 167 m, the resultant subaerial ice cliff height at the point of flotation being 14 m. Under such circumstances even minor changes to ice overburden pressure or water depth on the lower glacier would result in the flotation of the terminus (Benn et al., 2007). Flotation at freshwater termini is thought to occur if bending stresses are applied over a significant length of time for internal deformation to accommodate floatation (Warren et al., 2001; Boyce et al., 2007). Once at the point of flotation, calving events and terminus breakup result from relatively small, rapidly applied perturbations in proglacial water level (O'Neel et al., 2003; Boyce et al., 2007). Such perturbations at freshwater calving margins are caused by, for example, increased meltwater discharge or rainfall. Slow perturbations at freshwater calving margins may therefore allow for viscoelastic deformation rather that calving failure (Boyce, 2006). Boyce et al. (2007) demonstrated this with the example of Mendenhall Glacier, Alaska, where flotation of a section of the glacier was sustained over a two-year period by ice creep and lateral support of grounded ice near bedrock margins. A similar process is likely to have occurred at Tasman Glacier as it approached flotation due to decreased ice overburden pressure from supraglacial pond growth and retreat into deeper water. The relatively rapid retreat and disintegration of the MIC was therefore inevitable as the glacier retreated up-valley into a deeper part of the glacially eroded basin, at which point any significant changes in the stress regime on the MIC would cause the terminus to disintegrate.

RAINFALL AND TERMINUS DISINTEGRATION

Supraglacial pond fluctuations and buoyancy may also be linked to rainfall variability. According to the Mt Cook village NIWA climate station 7 km from the Tasman Glacier terminus (NIWA, 2008), mean monthly rainfall between 2000 and 2008 was 302 mm (Fig. 6). Maximum monthly rainfall occurred during November 2006 when 789 mm fell, which was preceded by 542 mm of rainfall in October 2006. This immediately preceded the large-scale retreat of the MIC terminus position after November 2006. Two factors linked to the high rainfall may have been important: (1) routing of meltwater and rainfall into supraglacial ponds increasing temperatures and water circulation within hydraulically connected ponds enhancing pond expansion by subaerial and subaqueous calving (Röhl, 2008); and (2) an increase in lake level. An increase in lake level of c. 4 m was recorded by Röhl (2005) in response to high rainfall, and the effect of water level changes have been documented at other calving glaciers (e.g. Laumann and Wold, 1992; Kirkbride and Warren, 1997; Motyka et al., 2003; Boyce et al., 2007), where the increase in lake level effects the flotation thickness of the glacier.

A secondary effect of this high rainfall event may have been an increase in glacier velocity associated with changes in the hydrological system of Tasman Glacier. At nearby Maud Glacier (Fig. 1) increases in glacier velocity associated with changes in hydrological system caused by rainfall were recorded by Kirkbride and Warren (1997). Associated uplift of the glacier surface (c. 0.2 m) during periods of higher velocities indicate that effective pressure (the difference between local ice overburden and basal water pressure) at the base of the glacier approached zero during intense rainfall events (e.g. O'Neel et al., 2003). Kirkbride and Warren (1997) noted that no evidence for a change in calving rate during periods of rainfall-induced velocity increases, resulting in the authors concluding that short-term velocity variations had no impact on calving. Maud Glacier is a grounded freshwater glacier that attained a stable terminus position in the mid-1990s, and is currently well in excess of its flotation thickness (Kirkbride and Warren, 1997).

No continuous velocity data are available for Tasman Glacier during the period of terminus disintegration, although from the example of Maud Glacier it is conceivable that a minor increase in velocity associated with a decrease in effective basal water pressure, caused by intense rainfall, would have altered the stress regime at the Tasman Glacier terminus. Thus, the high rainfall concurrent with the culmination of a period of rapid supraglacial pond growth and the associated increase in lake level would have enhanced any buoyancy-induced torque at the glacier terminus (Warren et al., 2001). Crack propagation at the glacier bed (e.g. Van der Veen, 1998) would have been initiated during this period of increased buoyancy, resulting in the disintegration of the terminus (Fig. 4). Evidence for this type of calving during the period of retreat is shown in Figure 4, where large (coherent) tabular icebergs that have calved from the terminus of Tasman Glacier can be seen moving down the lake. Such icebergs are similar to those described by Warren et al. (2001) at Glacier Nef, Patagonia, highlighting their origin as calved due to buoyancy of the terminus. Hence, rather than subaerial calving being the primary process of retreat at the MIC at the present time, large-scale buoyancy-driven calving is the major cause of contemporary terminus retreat.

ICE RAMP DEVELOPMENT

As highlighted by Quincey and Glasser (2009), establishing the presence or absence of a projecting ramp of ice at calving glacier termini may be a significant step toward more accurate mechanics-based prediction of future calving retreat. In particular,

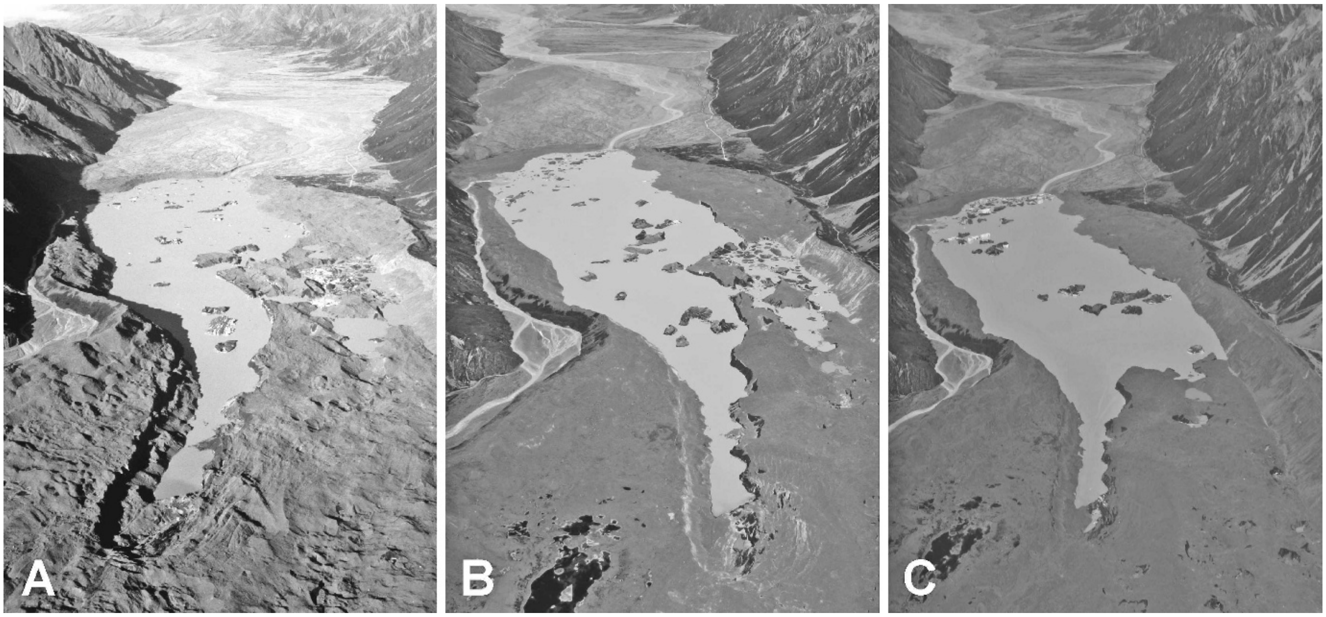

FIGURE 4. Oblique aerial photographs showing the disintegration of the lower Tasman Glacier terminus between (A) March 2006, (B) 2007, and (C) 2008 (photos: S. Winkler).

whether or not ice ramp development is a self-limiting process within a broader calving cycle is intriguing (e.g. Benn et al., 2007).

The precise nature and importance of ice ramps of freshwater-calving glaciers is poorly understood. Warren and Kirkbride (1998) identified the presence of small 1- to 2-m-wide ice ramps just below the waterline at Maud, Godley, and Hooker lakes in *Aoraki*/Mt Cook National Park (Fig. 1). However, from witnessing the emergence of icebergs <200 m from the calving terminus at Maud Glacier, they indicated that a substantial glacier-wide low-gradient ice ramp may have been present. At Tasman Glacier, Hochstein et al. (1995) and Purdie and Fitzharris (1999) indicated that the terminal ice cliff was almost vertical, while from observing stones and boulders falling from the glacier surface into Tasman Lake, Warren and Kirkbride (1998)

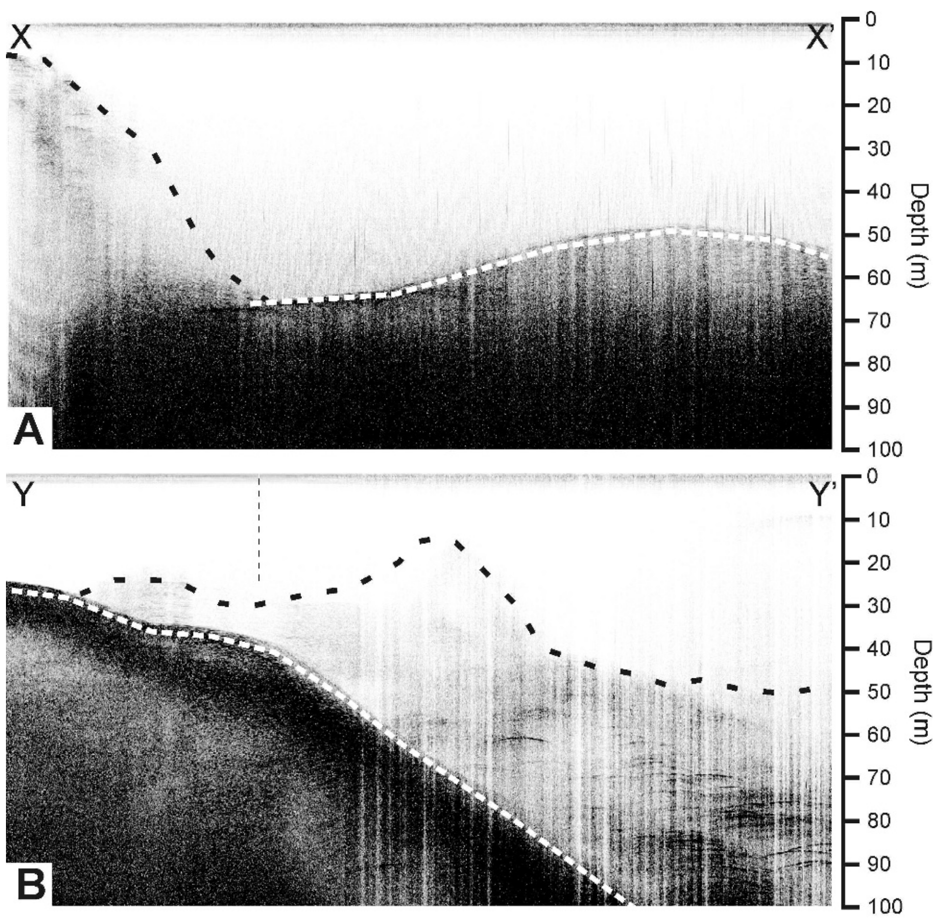

FIGURE 5. Sub-bottom profiles running perpendicular (A) and parallel (B) to the 2008 main ice cliff. In (A) the ice ramp (black dotted line) joins the bed of the lake (white dotted line), whereas in (B) the ice ramp is overlying the moraine wall (black dotted line) at the northeastern margin of the lake. See Figure 2 for profile locations.

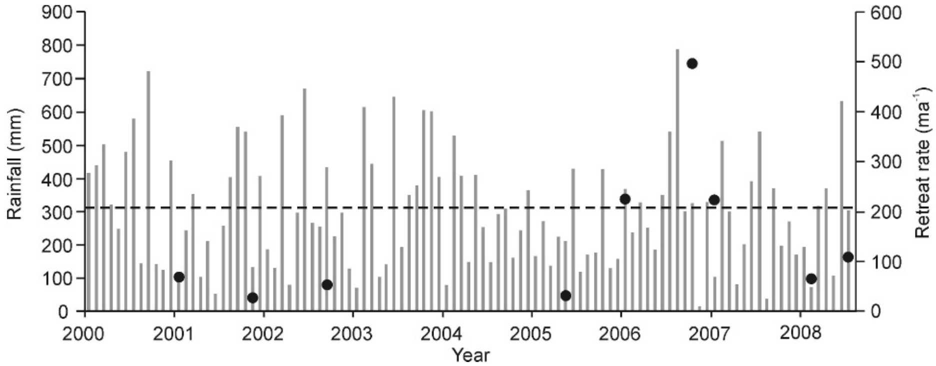

FIGURE 6. Monthly rainfall at Mount Cook Village (bars) and full width retreat rate of Tasman Glacier (circles) for the period 2000–2008. Dotted line indicates mean precipitation between 2000 and 2008. Date from NIWA (2008).

suggested the presence of a projecting ice ramp <4 m from the ice cliff. Our results (Fig. 5) indicate that a major >100-m-wide ice ramp with a complex surface does project out into the lake from across the glacier terminus.

Our observations of a subaqueous ice ramp at Tasman Glacier are inconsistent with the findings of Hochstein et al. (1995), who indicated that the submerged part of the glacier terminus was near-vertical, with no projecting ramp, standing in water almost 120 m deep. Our data show subaqueous ice ramping at large distances (>200 m) from the calving terminus, and this inconsistency is intriguing. One explanation for this discrepancy is that the data reported by Hochstein et al. (1995) are not representative of the majority of the Tasman calving terminus, and that a subaqueous ice ramp projecting away from the terminus was generally present in the mid-1990s. An alternative explanation is that the lake has become more thermally stratified as it has evolved since the mid-1990s. Hochstein et al. (1995) reported that Tasman Lake was almost isothermal, with a temperature range of 0.2 °C, and generally around ≤0.5 °C. However, more recent observations by Röhl (2006) indicate that lake limnology has changed, and water temperatures now probably aid ice ramp development. Indeed, in Röhl's (2006) study, temperatures ranged from 3 to 10 °C at the surface, dropping to under 1 °C below 40 m depth (Röhl, 2006, Fig. 8). This thermal stratification would enhance retreat toward the surface, and minimize melting at depth.

Conclusion

Whilst a period of terminus extension occurred at short (<10 year) response-time New Zealand glaciers during the 1980s and 1990s due to precipitation-driven positive mass balances, the longer response-time, debris-mantled, low-gradient glaciers such as Tasman Glacier commenced a phase of calving retreat. This study has documented the retreat of Tasman Glacier and the evolution of Tasman Lake, utilizing multispectral satellite imagery for the period 2000–2008. The retreat of Tasman Glacier has been rapid compared to similar water-terminating glaciers in the *Aoraki*/Mt Cook National Park, although significant temporal and spatial variations in the rate of retreat have been identified. Temporally, there is an identifiable transition in the rate of retreat prior to, and after, 2006. Spatially, retreat of the glacier terminus during 2000–2006 can be categorized into the rapidly retreating main ice cliff (MIC) and the comparatively slowly retreating eastern embayment ice cliff (EEIC). Retreat prior to 2006 was characterized by low magnitude–high frequency iceberg calving, 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 accurately deconvolve the controlling mechanisms from a retrospective study, evidence suggests buoyancy-driven calving became the primary mechanism of calving for the MIC. A combination of supraglacial pond growth, rainfall, and water-depth increase at the terminus as the glacier retreated up-valley may have led to increased buoyancy, initiating crack propagation at the bed and triggering terminus disintegration (e.g., Benn et al., 2007). Limits to future progression of calving retreat at Tasman Glacier are open to conjecture. At 10 km from the LIA terminus, the glacier bed is c. 350 m below the surface level of Tasman Lake, and 20 km upstream from the LIA terminus, the glacier bed lies at c. 950 m (Broadbent, 1974). Interpolation between these points indicates a further 8–10 km of calving retreat may be possible, perhaps to around the confluence with the tributary Hochstetter Glacier. The presence of a substantial projecting ice ramp at the terminus of Tasman Glacier is intriguing, and such features are likely to have an influence on the future retreat rates at this and other lake-calving glaciers in the area. The effects of future atmospheric circulation variability on the mass balance of New Zealand glaciers are uncertain. Chinn et al. (2005) described how glacier expansion may occur in the Southern Alps even during periods of small increases in surface air temperature, if atmospheric circulation produces marked increase in precipitation. However, the continuing calving retreat of Tasman Glacier demonstrates that glaciers also respond to non-climatic influences. Therefore, the precise geometry of ice ramps, the degree to which they affect subaerial ablation processes, and the effect of ramps on future subaqueous calving retreat and lake expansion are clear areas for further research.

Acknowledgments

Stefan Winkler is thanked for providing photographs and comments on earlier versions of this work. The fieldwork was funded by Massey University Research Fund (MURF), and Ray Bellringer from the Department of Conservation is thanked for permitting fieldwork in the National Park. David Feek's expertise in the boat was extremely valuable, as was John Appleby's field assistance. Mike Tuohy kindly provided assistance with satellite imagery. We thank the associate editor and two reviewers for helping to considerably improve this manuscript.

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MS accepted July 2010