Annals of Glaciology 36 2003

Calving speed and climatic sensitivity of New Zealand lake-calving glaciers

Charles R. WARREN,[1] Martin P. KIRKBRIDE[2]
[1]*School of Geography and Geosciences, University of St Andrews, St Andrews, Fife KY16 9AL, Scotland*
E-mail: charles.warren@st-andrews.ac.uk
[2]*Department of Geography, University of Dundee, Dundee DD1 4HN, Scotland*

ABSTRACT. Calving speeds and calving mechanisms in fresh water contrast with those in tidewater. We obtained calving speeds for six lake-calving glaciers in New Zealand's Southern Alps, and surveyed the depths and temperatures of their ice-contact lakes. The glaciers are temperate, grounded in shallow ($\leq$20 m) water, and exhibit compressive flow at their termini. These data increase the global dataset of fresh-water calving statistics by 40%, bringing the total to 21 glaciers. For this dataset, calving rates (u_c) correlate positively with water depths (h_w) ($r^2 = 0.83$), the relationship being expressed by: $u_c = 17.4 + 2.3h_w$. This is an order of magnitude lower than values of u_c at temperate tidewater glaciers. For a subset of 10 glaciers for which ice-proximal water temperature (t_w) data are available, u_c also correlates positively with t_w, supporting a physical relation between calving and melting at and below the water-line. Fluctuations of New Zealand lake-calving glaciers in the period 1958–97 show that although the transition from non-calving to calving dramatically increases frontal retreat rates, the onset of calving does not isolate terminus change from climatic forcing. In terms of climatic sensitivity, lake-calving glaciers occupy an intermediate position between tidewater glaciers (least sensitive) and non-calving glaciers (most sensitive).

CALVING IN FRESH WATER

This paper examines the calving rates and climatic sensitivity of lake-calving glaciers in New Zealand's Southern Alps. The aims are (1) to extend the database of calving statistics for lake-calving glaciers, and (2) to establish the glaciological and glacio-climatic significance of the non-calving/calving transition. Most calving research has focused on tidewater glaciers (Van der Veen, 1997, 2002). Lake-calving glaciers have received much less attention, despite the strong contrast in calving mechanisms and rates between tidewater and fresh-water settings (Warren and others, 1995; Skvarca and others, 2002), and the fact that lake-calving glaciers can represent significant hazards and resources. Possible reasons for the different calving rates include the contrasting buoyancy of meltwater in fresh and salt water, contrasts in water chemistry and longitudinal strain rates, and differences in flow speeds resulting from buoyancy effects at the terminus (Funk and Röthlisberger, 1989; Warren and others, 1995; Van der Veen, 2002), but empirical testing of these possibilities is hampered by the dangerous environment. Theoretical and modelling approaches (Hughes, 1992, 2002; Hanson and Hooke, 2000; Vieli and others, 2001) bypass this difficulty, but their generic nature makes their specific applicability uncertain. Given the site-specific nature of calving processes, Van der Veen (1997) identifies the building of regional databases of calving statistics as a key research priority. This work constitutes a response to that need.

The potential for calving glaciers to respond indirectly to climate forcing on decadal to millennial time-scales is well established (Meier, 1997; Trabant and others, 2003). However, the bipolar scheme in which non-calving glaciers are represented as climatically sensitive, and calving glaciers as insensitive, is challenged by Warren (1991) who proposes a tripartite division, differentiating between tidewater and fresh-water glaciers. He suggests a continuum of increasing climatic sensitivity from tidewater through fresh-water to non-calving glaciers. This study tests this idea through an examination of the recent fluctuation histories of lake-calving glaciers in New Zealand (where no glaciers reach tidewater).

DATA COLLECTION

Seven calving glaciers were studied during the austral autumns of 1994 and 1995, and six were revisited and/or overflown in July 1997. All are temperate, grounded, debris-covered and largely uncrevassed. They terminate in turbid ice-contact lakes at calving faces typically 25–40 m high. The lakes have warm surface layers in summer but freeze over in winter, resulting in minimal calving and ablation (Warren and Kirkbride, 1998). At Grey and Maud Glaciers, annual velocities were derived from electronic-theodolite survey of marked supraglacial boulders. Short-term (7–19 day) surface ice velocities were derived from electronic-theodolite surveys at Maud, Grey, Godley and Ruth Glaciers in 1994, and at Maud, Grey, Tasman and Hooker Glaciers in 1995, using transverse and longitudinal profiles of surface markers. The transverse profiles comprised 5–9 points distributed across the full width of each glacier and were located a few metres to decimetres up-glacier from the calving cliffs. The longitudinal profiles (2–4 points) were located close to the geometric centre lines. Errors cannot be quantified from single-point

https://doi.org/10.3189/172756403781816446 Published online by Cambridge University Press

Table 1. Calving data for the study glaciers, 1994–95

Glacier	h_w	t_w	u_i	dL/dt	u_c	$u_c + u_m$
	m	°C	m a^{-1}	m a^{-1}	m a^{-1}	m a^{-1}
Maud	15	4.3	151	−45	88	106
Grey	12	4.2	52	13	47	65
Godley	17	3.1	5	60	47	65
Ruth	4	3.1	6	30	18	36
Tasman	10	0.5	11	35	28	46
Hooker	20	1.7	28	4	14	32

Notes: h_w is water depth, t_w is water temperature at 10 m depth, u_i is ice velocity, dL/dt is retreat rate, u_c is calving rate and u_m is melt rate. dL/dt, u_i and u_c are all width- and annually averaged, and h_w is width-averaged. For Hooker Glacier, dL/dt is for the period 1995–97. t_w is from Warren and Kirkbride (1998) except for Tasman Glacier for which t_w is from Hochstein and others (1995). We have inferred that sloping proglacial lake floors consist of debris-covered glacier ice (Warren and Kirkbride, 1998). If so, these ice-proximal values of h_w are effective depths that are not indicative of ice thickness.

surveys, but are estimated to be ≤1% of the annual velocities. At Maud Glacier, both aerial and terrestrial photogrammetry were also employed to determine the surface velocity field of the lower 400 m of the glacier (Kirkbride and Warren, 1997). Rates of terminus change and ice speeds were width-averaged for 1994–95 except for Hooker Glacier at which terminus change is for the period 1995–97.

Most calculations of calving speed implicitly include melting of ice cliffs, on the assumption that melt rates are negligible compared to calving losses, but the mid-latitude location and slow calving speeds of these glaciers demand that melting be considered separately. Ablation was measured at six stakes at Maud and Grey Glaciers in 1994; four were placed in steep (45–75°) ice walls close to the termini, one on the surface of Grey Glacier and one in the true-right end of the vertical frontal cliff of Maud Glacier (cf. Kirkbride and Warren, 1997). Measurements adopted the straight-edge method and were accurate to ±15 mm. The bathymetry and vertical water-temperature structure of the ice-contact lakes at Hooker, Maud–Grey and Godley–Ruth Glaciers were surveyed using methods described by Warren and Kirkbride (1998). Patterns of frontal change were constructed using vertical aerial photographs dating from 1958, 1965, 1971, 1974 and 1985/86. Frontal positions in 1994, 1995 and 1997 were derived from ground survey and from near-vertical and oblique aerial photographs taken for the purpose. Classen Glacier was included in the frontal-change analysis but not in the field programme.

RESULTS

Calving rates

Calving data are presented in Table 1. Calving rates (u_c in m a^{-1}) were obtained from the equation:

$$u_c = u_i - \frac{dL}{dt} - u_m \,, \tag{1}$$

where u_i is ice velocity, dL/dt is rate of change in glacier length, and u_m is melt rate at the calving face, all in m a^{-1} and width-averaged. Because u_c is low, melting accounts for significant proportions of total retreat (17–56%), but u_m is the least well-constrained term in Equation (1). Though limited by site constraints, our ablation data yield a mean cliff-melt rate of 50 mm d^{-1} over 11 days. During more rigorous measurements on Tasman Glacier from 1985 to 1987 (Kirkbride, 1995) an ablation rate of 47 mm d^{-1} was measured over 28 days in the same season and at a similar altitude. The similarity of these values suggests that both the magnitude and seasonal variability of ablation rates at the study glaciers (which terminate at altitudes of 727–1095 m with calving cliffs orientated southwest to southeast) are comparable. This permits the annual contribution of melt to glacier retreat to be estimated from our short-term measurements, yielding a value of 18.0 m a^{-1} (cf. the measured annual value of 17.7 m a^{-1} at Tasman Glacier (Kirkbride, 1995)). This single value of u_m has then been applied at all sites to provide estimates of rates of mechanical ice loss. Since most published calving rates disregard the melt term, we used $u_c + u_m$ for the New Zealand glaciers when testing the dependence of u_c on water depth (h_w).

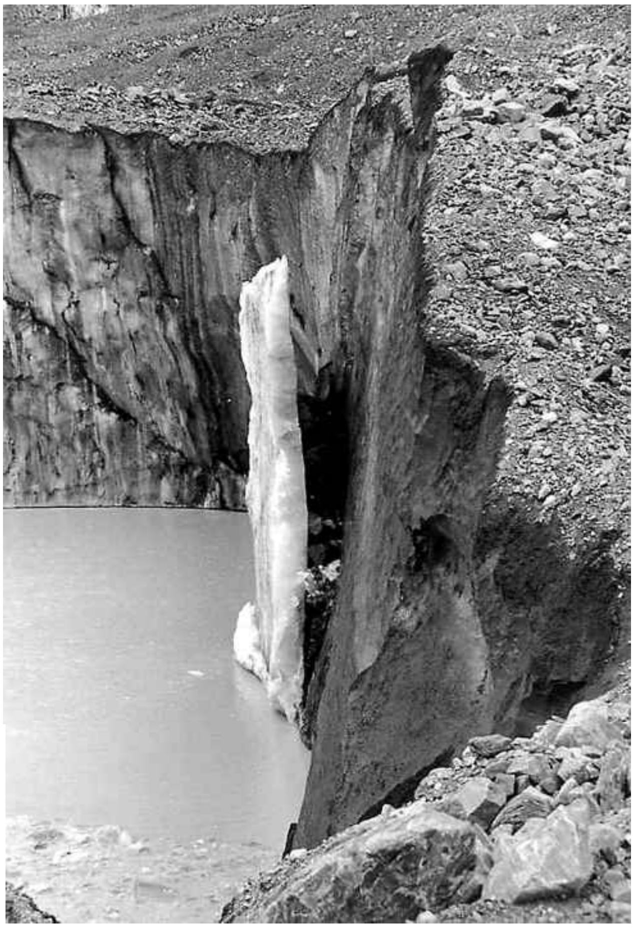

Fig. 1. Godley Glacier terminus, 2 April 1994: thin flake prior to calving. The flake measured about 18 m high by 14 m wide by 1 m thick. The calving cliff at this point rises 26 m above lake level.

Regression of h_w against u_c was carried out for (1) the global dataset compiled by Warren and others (1995), and (2) these data with the addition of eight further glaciers: the six New Zealand glaciers; Svartisheibreen, Norway (Kennett and others, 1997); and Glaciar Ameghino, Argentina (Warren, 1999) — a total of 21 glaciers. Coefficients of variation (r^2) are 0.75 and 0.83, respectively, significant at the 0.01 level. Correlations using maximum water depth instead of h_w are slightly weaker. The global relationship is now described by:

$$u_c = 17.4 + 2.3h_w \,. \tag{2}$$

https://doi.org/10.3189/172756403781816446 Published online by Cambridge University Press

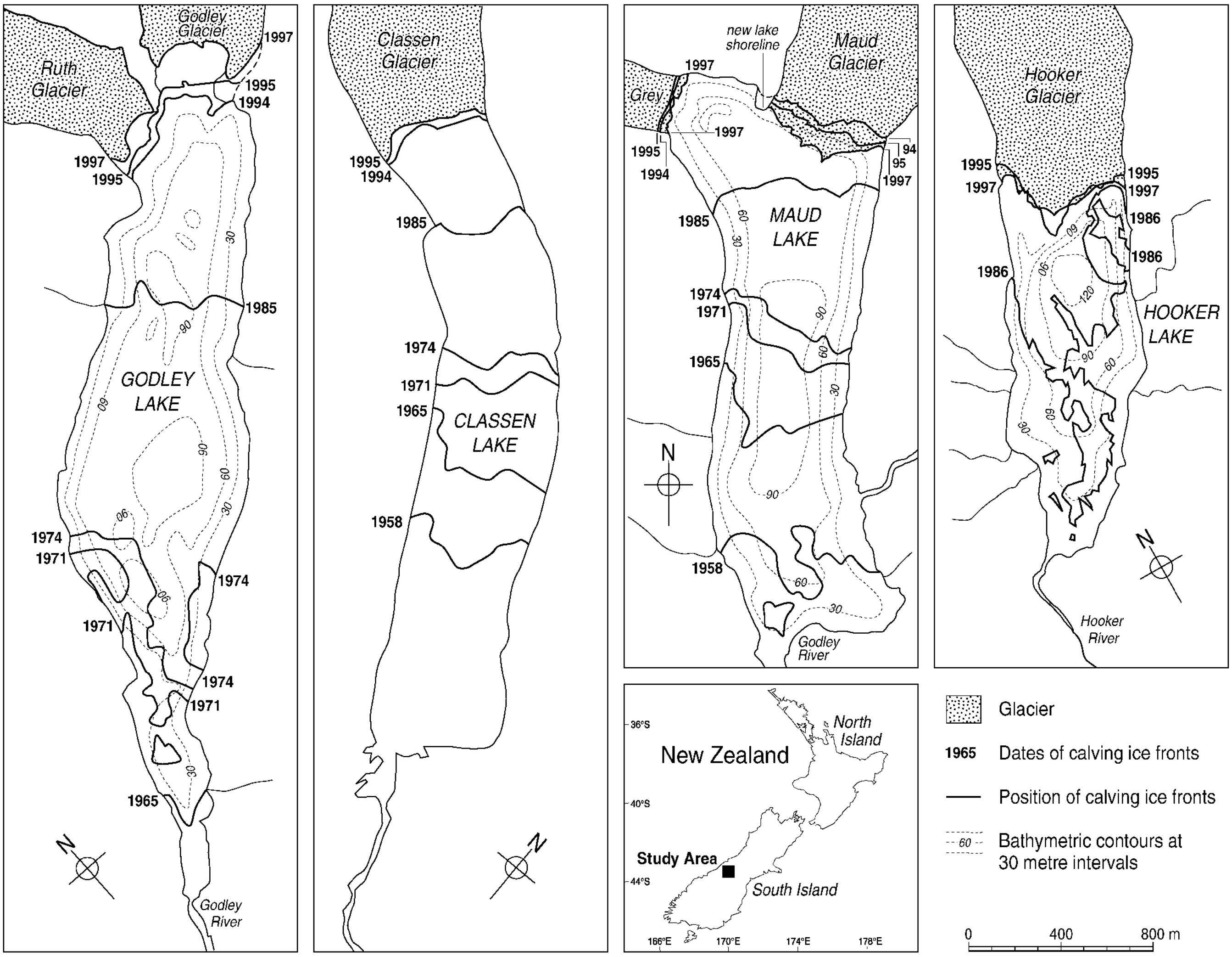

Fig. 2. Fluctuation histories for six of the study glaciers. Bathymetry is simplified from Warren and Kirkbride (1998); bathymetric data do not exist for Classen Lake. In 1965, Hooker Glacier filled the current lake basin, and lake initiation began in the 1970s. The fluctuation history of Tasman Glacier, also studied during this research, is discussed by Kirkbride and Warren (1999).

This applies for grounded lake-calving termini; data from sites where calving is driven by buoyant forces at floating or near-floating termini (e.g. Glaciar Nef, Chile (Warren and others, 2001); Glaciar Upsala, Argentina (Skvarca and others, 2002)) have been excluded. Calving at all the New Zealand glaciers occurs through the four-stage mechanism of thermal undercutting described in detail by Kirkbride and Warren (1997). This mechanism is driven by thermo-erosion at the water-line followed by the progressive development of an overhang through flake calving (or spalling) from the subaerial cliff face. Such flakes often have high width/thickness ratios, as high as 25 : 1 (Fig. 1). Sub-horizontal melt notches at the water-line typically extend $\leq$4 m into the base of the cliff, but Purdie and Fitzharris (1999) report undercuts up to 10 m deep at Tasman Glacier. Retreat of the subaerial cliff produces a subaqueous ice foot from which all the largest icebergs are calved.

Fluctuation histories

Fluctuation histories and areal retreat rates are shown in Figures 2 and 3. Given the highly irregular patterns of retreat during lake initiation, areal rates of change are more meaningful than linear rates. During the early stages of lake formation, when low-gradient termini are gradually inundated, retreat rates increase only slightly, but the onset of calving at terminal cliffs dramatically increases retreat rates (e.g. the 19-fold acceleration at Hooker Glacier, 1982–86 (Fig. 3)). At Grey–Maud and Classen Glaciers, this acceleration occurred pre-1958 and is documented by Kirkbride (1993); at Godley and Grey–Maud Glaciers, recession increased by factors of 10 and 4, respectively. These transitions coincided with the period of greatest 20th-century warming, when retreat rates would have increased without lake formation (Chinn, 1996), but much of the acceleration can nevertheless be attributed to the onset of calving. Despite this shift into a more negative net mass-balance regime, a climate signal can still be discerned in the fluctuations of the calving termini. This does not apply during the three-stage dynamic adjustment from non-calving to calving (cf. Kirkbride, 1993) (e.g. Godley Glacier, 1965–74). However, once a calving cliff (as opposed to a drowned terminus) has been established, a measure of synchrony in terminus behaviour is apparent. For example, apparently synchronous reductions in retreat rate occur in the early 1970s at Grey–Maud and Classen Glaciers, and retreat rates increased at all glaciers after 1974. Most strikingly, readvances and/or reductions in retreat rate took place at all sites in the mid-1990s.

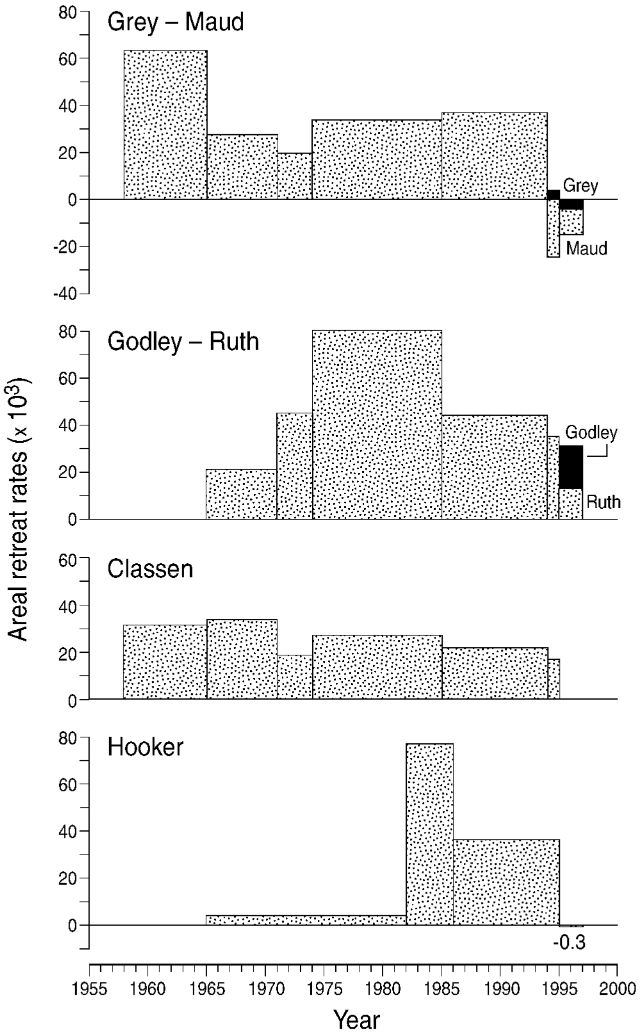

Fig. 3. Areal retreat rates of the study glaciers, 1958–97. Grey–Maud and Godley–Ruth were confluent prior to 1990 and 1996, respectively. For Hooker Glacier, a 1982 lake area of 70 000 m^2 has been taken from Hochstein and others (1998).

DISCUSSION

Calving in fresh water

A linear correlation between u_c and h_w exists in a diversity of geographical settings, but the constant of proportionality varies widely. In particular, existing data consistently show that calving occurs much more slowly in lakes than in comparable tidewater settings. This applies for grounded termini in shallow and deep water (Warren and others, 1995) and for calving fronts which reach flotation (Warren and others, 2001; Skvarca and others, 2002). Van der Veen (2002) has challenged the water-depth model, arguing that the u_c/h_w correlation does not signify causality. Hanson and Hooke (2000), however, advance some possible physical explanations for the observed u_c/h_w relation. Their modelling of grounded tidewater fronts suggests that the development of flow-induced oversteepening of calving faces, possibly exacerbated by submarine melting, may partly explain the correlation. Causative or not, the New Zealand data confirm that u_c correlates linearly with h_w in fresh water, and that, for any given water depth, grounded temperate glaciers calve more than an order of magnitude slower in fresh water than in tidewater. These data also suggest that, at least for temperate glaciers, calving rates and mechanisms are driven by the effects of limnological processes on glaciological processes.

At cold-based glaciers, Iken (1977) and Hughes (1992, 2002) have suggested that the rate-controlling process is forward-bending along englacial shear bands created by unrelieved tensile stresses. At grounded, temperate lake-calving glaciers we have proposed that water-line melting is the rate-controlling process (Kirkbride and Warren, 1997). Purdie and Fitzharris (1999) also document a causal relationship between thermo-erosional water-line notches and subaerial calving. Overhangs are common along subaerial sections of fresh-water termini, but these are created by thermal undercutting and progressive spalling above thermo-erosional notches, not primarily by forward bending. At compressional termini, full-height slab calving is defined by surface crevasses which open in response to the combination of water-line melting and increasing overhang of the subaerial cliff. Slab calving occurs too frequently to allow overhangs to be created by the bending creep mechanism proposed by Hughes (1992, 2002). This and other arguments have led some (e.g. Hanson and Hooke, 2000) to suggest that Hughes' elastic-beam theory approach is inappropriate. Similarly, we found that glacier dynamics more than a few metres upstream from the cliffs did not affect, nor were affected by, the calving process (Kirkbride and Warren, 1997). Our observations do not rule out the possibility that oversteepenings are partly created by the distribution of longitudinal stresses and velocities, as proposed by Hanson and Hooke (2000), but they show that overhangs can form primarily in response to water-line melting.

The geometry of the submerged parts of calving termini has long been debated (cf. discussion in Warren and Kirkbride, 1998). The existence of projecting platforms or "ice feet", inferred from the distal emergence of subaqueous icebergs, has been rejected by some in favour of vertical or even undercut submerged ice cliffs, a geometry also inferred for Tasman and Hooker Glaciers by Hochstein and others (1995, 1998). However, our observations and those of Purdie and Fitzharris (1999) show that ice feet are present along the majority of these termini, commonly consisting of a small (1–4 m) sub-horizontal shelf just below the water-line and a ramp of ice dipping lakewards at 20–45°. Ramps may intersect the cliff close to the water-line or at depth. Ice feet have also been reported at Alaskan tidewater glaciers (Motyka, 1997; Hunter and Powell, 1998) and may be significant through their control on effective h_w (cf. Table 1 caption), circulation patterns and hence heat advection to calving fronts.

The glaciological results reported here represent an end-member of the spectrum of calving behaviour, the other extreme being catastrophically retreating tidewater glaciers such as Columbia Glacier, Alaska, U.S.A., at which calving rates are two orders of magnitude greater (Pfeffer and others, 2000). The styles of calving observed in New Zealand have not been reported at temperate tidewater sites, presumably because the frequency of calving exceeds the rate of melt-water notch development and because calving in such settings is defined by crevasses formed up-glacier of the termini by extending flow. At glacier termini where velocities are low and crevassing is limited because of compressive flow, fresh-water calving is driven by water-line melting. This may also be the case at polar tidewater glaciers (A. Vieli and others, unpublished information). Aquatic factors appear to dominate at compressional margins, and glaciological factors at extensional margins.

https://doi.org/10.3189/172756403781816446 Published online by Cambridge University Press

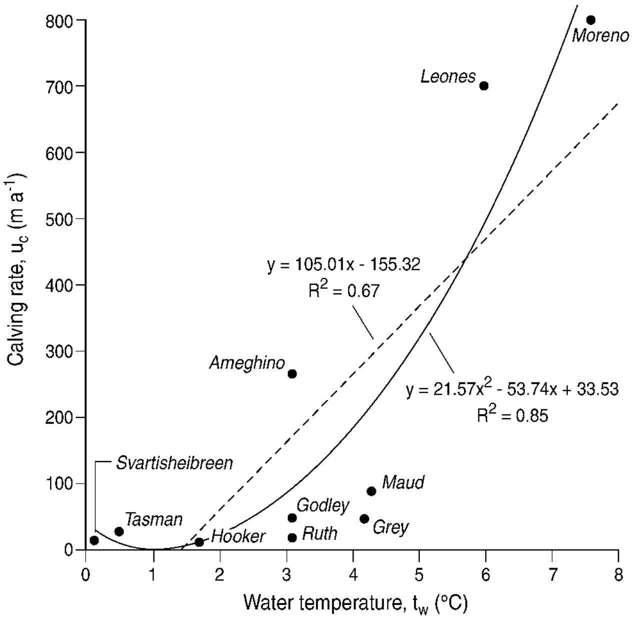

Fig. 4. Calving rates plotted against water temperatures for New Zealand glaciers in this study, and for the following glaciers: Glaciar Moreno, Argentina (Rott and others, 1998; Warren, 1999); Glaciar Ameghino, Argentina (Warren, 1999); Glaciar Leones, Chile (E. Haresign and C. R. Warren, unpublished); and Svartisheibreen, Norway (Kennett and others, 1997). Ice-proximal water temperatures were measured in summer or early autumn at a depth of 10 m.

Climatic response of calving glaciers

The terminus fluctuations of all types of calving glaciers may at times be subject to non-climatic (notably topographic) controls (Warren, 1992; Meier, 1997). However, our data support the hypothesis that, in terms of climatic sensitivity, lake-calving glaciers occupy an intermediate position between tidewater glaciers (least sensitive) and non-calving glaciers (most sensitive). Glaciers in the Southern Alps thinned and retreated through much of the 20th century (Chinn, 1996). All the large valley glaciers retreated from terminal moraines and outwash heads, initiating lake formation (Kirkbride, 1993; Kirkbride and Warren, 1999), and the subsequent onset of calving greatly increased retreat rates, from a mean of 12 m a^{-1} prior to lake formation to a mean of 50 m a^{-1} thereafter (Chinn, 1996; Purdie and Fitzharris, 1999). Because of exceptionally heavy precipitation, New Zealand glaciers are at the high end of the glacier sensitivity range, responsive to small climate variations (Fitzharris and others, 1997). Although the onset of calving increases the rate of frontal recession, reducing this sensitivity, terminus changes do not become entirely isolated from climatically induced changes in mass balance. Periods of regionally positive mass balance, recorded by advances of non-calving glaciers, can be discerned as near-synchronous periods of readvances and/or reduced retreat rates of calving glaciers. Thus, for example, the readvances and reduced retreat rates observed in the mid-1990s (Fig. 3) probably represent a delayed response to the strongly positive mass-balance years of the 1980s and early 1990s which caused thickening and advance of most non-calving glaciers and lowering snowlines on calving glaciers (Chinn, 1996). A shift towards more positive mass balance increases ice velocities at the terminus and reduces the excess of calving losses over ice supply, thereby reducing retreat rates or (if sufficient) triggering readvances.

As would be expected for calving glaciers, the picture is neither straightforward nor uniform, partly because the glaciers underwent dynamic adjustment from non-calving to calving at different times, and partly because of the variable strength of topographic controls which are of great importance for calving-glacier fluctuations (Warren, 1991; Vieli and others, 2001). Nevertheless, some of the observed retreat-rate changes occur at locations of invariant topographic geometry and so cannot be attributed to topographically driven changes in mass balance. These complexities make it all the more striking that any measure of synchrony is apparent. It suggests that the onset of calving does not decouple these glaciers from climate, as it typically does in tidewater settings. Rather, it translates them into a new glacio-climatic regime of enhanced ablation which persists throughout the period of retreat through the lake basin. In this regime, annual net balances are consistently more negative than prior to the non-calving/calving transition, but changing climatic stimuli can still be discerned in patterns of frontal change. The reason for this may be that water-line melting is the rate-controlling mechanism. Since u_m at the water-line is likely to vary within a restricted range, u_c too may vary little during retreat through the lake basin, permitting climatically driven increases in ice supply to be expressed as fluctuating retreat rates.

However, if u_c is controlled by water-line melting, why does the u_c/h_w correlation persist? It is not clear whether the correlation indicates causation, or whether deeper water enhances some other process. For example, more vigorous lake circulation may enhance heat advection to the waterline, or faster subaqueous calving and/or melting may affect ice-cliff stability. Both possibilities suggest that ice-proximal water temperatures and circulation patterns may significantly affect calving rates and terminus dynamics, as suggested by Hanson and Hooke (2000) and Motyka and others (2003). Empirical data with which to test this suggestion are limited, but Figure 4 shows that u_c correlates positively with t_w. (All water temperatures were measured in the summer or early autumn within a few tens of metres of glacier termini and at a depth of 10 m.) The correlation is stronger with a non-linear than a linear fit ($r^2 = 0.85$ and 0.67, respectively, both significant at the 0.01 level), and the strength of correlation is similar to that between u_c and h_w in fresh water. Brown and others (1982) believe that there is no direct evidence that t_w needs to be considered explicitly as a separate variable at Alaskan tidewater glaciers, and propose a simple one-coefficient, one-independent-variable calving relation. In fresh water, the strength of the positive correlations between u_c, on the one hand, and both h_w and t_w, on the other, suggests that the correlation between u_c and h_w is largely explained by ice-proximal lacustrine processes.

The data in Figure 4 provide empirical support for Funk and Röthlisberger's (1989) suggestion that u_c is strongly dependent on water density through its impact on meltwater upwelling and subaqueous melt rates. They also support Hanson and Hooke's (2000) conclusion that h_w is probably only one of several variables affecting u_c, other likely candidates being water temperature and longitudinal strain rate. Van der Veen (2002) also appeals to water-density differences to explain the tidewater/fresh-water contrast in u_c, but suggests that they affect calving through their influence on effective basal pressures and hence sliding speeds near the terminus. The processes observed at slow, uncrevassed, lake-terminating glaciers such as those in New Zealand are likely to be overwhelmed by the much faster calving mechanisms which dominate at fast-flowing, heavily crevassed tidewater glaciers.

CONCLUSIONS

Our main conclusions are:

(1) Water-line and subaqueous processes are significant in explaining the contrast between calving rates in tidewater and fresh-water environments. A positive correlation between water temperature and calving rates suggests that water temperature may be a key variable alongside water depth in explaining calving speeds. In fresh water, this may be explained in terms of the thermal undercutting calving mechanism of Kirkbride and Warren (1997).

(2) Glaciers become progressively less climatically sensitive along the spectrum from slow to fast calving. The non-calving/calving transition causes a rapid increase in retreat rate, but even within this different mass-balance framework a climate signal can be discerned in the fluctuation histories of lake-calving glaciers.

(3) This work has augmented the global dataset of fresh-water calving statistics by 40%. The results strengthen the significance of the u_c/h_w correlation and confirm that, for grounded temperate glaciers, calving rates in fresh water are slower by a factor of ~11 than those in tidewater.

ACKNOWLEDGEMENTS

The work was funded by the U.K. Natural Environment Research Council (1994–95) and the Royal Society (1997). Field assistance from N. Brazier, S. Warren and K. Coombes is gratefully acknowledged. Permission to carry out the research was granted by Mount Cook National Park staff. Constructive refereeing by T. Hughes and R. Motyka improved the paper.

REFERENCES

Brown, C. S., M. F. Meier and A. Post. 1982. Calving speed of Alaska tidewater glaciers, with application to Columbia Glacier. *U.S. Geol. Surv. Prof. Pap.* 1258-C.

Chinn, T. J. 1996. New Zealand glacier responses to climate change of the past century. *N.Z. J. Geol. Geophys.*, **39**(3), 415–428.

Fitzharris, B. B., T. J. Chinn and G. N. Lamont. 1997. Glacier balance fluctuations and atmospheric circulation patterns over the Southern Alps, New Zealand. *Int. J. Climatol.*, **17**(7), 745–763.

Funk, M. and H. Röthlisberger. 1989. Forecasting the effects of a planned reservoir which will partially flood the tongue of Unteraargletscher in Switzerland. *Ann. Glaciol.*, **13**, 76–81.

Hanson, B. and R. LeB. Hooke. 2000. Glacier calving: a numerical model of forces in the calving-speed/water-depth relation. *J. Glaciol.*, **46**(153), 188–196.

Hochstein, M. P., D. Claridge, S. A. Henrys, A. Pyne, D. C. Nobes and S. F. Leary. 1995. Downwasting of the Tasman Glacier (South Island, N.Z.): changes in the terminus region between 1971 and 1993. *N.Z. J. Geol. Geophys.*, **38**(1), 1–16.

Hochstein, M. P., M. I. Watson, B. Malengrenau, D. C. Nobes and I. Owens. 1998. Rapid melting of the terminal section of the Hooker Glacier (Mt Cook National Park, New Zealand). *N.Z. J. Geol. Geophys.*, **41**(3), 203–218.

Hughes, T. 1992. Theoretical calving rates from glaciers along ice walls grounded in water of variable depths. *J. Glaciol.*, **38**(129), 282–294.

Hughes, T. J. 2002. Calving bays. *Quat. Sci. Rev.*, **21**(1–3), 267–282.

Hunter, L. E. and R. D. Powell. 1998. Ice foot development at temperate tidewater margins in Alaska. *Geophys. Res. Lett.*, **25**(11), 1923–1926.

Iken, A. 1977. Movement of a large ice mass before breaking off. *J. Glaciol.*, **19**(81), 595–605.

Kennett, M., T. Laumann and B. Kjøllmoen. 1997. Predicted response of the calving glacier Svartisheibreen, Norway, and outbursts from it, to future changes in climate and lake level. *Ann. Glaciol.*, **24**, 16–20.

Kirkbride, M. P. 1993. The temporal significance of transitions from melting to calving termini at glaciers in the central Southern Alps of New Zealand. *Holocene*, **3**(3), 232–240.

Kirkbride, M. P. 1995. Relationships between temperature and ablation on the Tasman Glacier, Mount Cook National Park, New Zealand. *N.Z. J. Geol. Geophys.*, **38**(1), 17–27.

Kirkbride, M. P. and C. R. Warren. 1997. Calving processes at a grounded ice cliff. *Ann. Glaciol.*, **24**, 116–121.

Kirkbride, M. P. and C. R. Warren. 1999. Tasman Glacier, New Zealand: 20th-century thinning and predicted calving retreat. *Global Planet. Change*, **22**(1–4), 11–28.

Meier, M. F. 1997. The iceberg discharge process: observations and inferences drawn from the study of Columbia Glacier. *Byrd Polar Res. Cent. Rep.* 15, 109–114.

Motyka, R. J. 1997. Deep-water calving at Le Conte Glacier, southeast Alaska. *Byrd Polar Res. Cent. Rep.* 15, 115–118.

Motyka, R. J., L. Hunter, K. A. Echelmeyer and C. Connor. 2003. Submarine melting at the terminus of a temperate tidewater glacier, LeConte Glacier, Alaska, U.S.A. *Ann. Glaciol.*, **36** (see paper in this volume).

Pfeffer, W. T., J. B. Cohn, M. F. Meier and R. M. Krimmel. 2000. Alaskan glacier beats a dramatic retreat. *Eos*, **81**(48), 577, 584.

Purdie, J. and B. Fitzharris. 1999. Processes and rates of ice loss at the terminus of Tasman Glacier, New Zealand. *Global Planet. Change*, **22**(1–4), 79–91.

Rott, H., M. Stuefer, A. Siegel, P. Skvarca and A. Eckstaller. 1998. Mass fluxes and dynamics of Moreno Glacier, Southern Patagonia Icefield. *Geophys. Res. Lett.*, **25**(9), 1407–1410.

Skvarca, P., H. de Angelis, R. Naruse, C. R. Warren and M. Aniya. 2002. Calving rates in fresh water: new data from southern Patagonia. *Ann. Glaciol.*, **34**, 379–384.

Trabant, D.C., R. M. Krimmel, K. A. Echelmeyer, S. Zirnheld and D. H. Elsberg. 2003. The slow advance of a calving glacier: Hubbard Glacier, Alaska, U.S.A. *Ann. Glaciol.*, **36** (see paper in this volume).

Van der Veen, C. J., *ed.* 1997. Calving glaciers: report of a Workshop, February 28–March 2, 1997, Columbus, OH. *Byrd Polar Res. Cent. Rep.* 15.

Van der Veen, C. J. 2002. Calving glaciers. *Prog. Phys. Geogr.*, **26**(1), 96–122.

Vieli, A., M. Funk and H. Blatter. 2001. Flow dynamics of tidewater glaciers: a numerical modelling approach. *J. Glaciol.*, **47**(159), 595–606.

Warren, C. R. 1991. Terminal environment, topographic control and fluctuations of West Greenland glaciers. *Boreas*, **20**(1), 1–15.

Warren, C. R. 1992. Iceberg calving and the glacioclimatic record. *Prog. Phys. Geogr.*, **16**(3), 253–282.

Warren, C. R. 1999. Calving speed in freshwater at Glaciar Ameghino, Patagonia. *Z. Gletscherkd. Glazialgeol.*, **35**(1), 21–34.

Warren, C. R. and M. P. Kirkbride. 1998. Temperature and bathymetry of ice-contact lakes in Mount Cook National Park, New Zealand. *N.Z. J. Geol. Geophys.*, **41**(2), 133–143.

Warren, C. R., D. R. Greene and N. F. Glasser. 1995. Glaciar Upsala, Patagonia: rapid calving retreat in fresh water. *Ann. Glaciol.*, **21**, 311–316.

Warren, C., D. Benn, V. Winchester and S. Harrison. 2001. Buoyancy-driven lacustrine calving, Glaciar Nef, Chilean Patagonia. *J. Glaciol.*, **47**(156), 135–146. (Erratum: **47**(157), p. 348.)