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Seasonal Variation in Ablation and Surface Velocity on a Temperate Maritime Glacier: Fox Glacier, New Zealand

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Abstract

Seasonal variations in ablation and surface velocity were investigated on the lower part of Fox Glacier, South Westland, New Zealand. A large variation between summer and winter ablation was recorded, with daily averages of 129 mm d^{-1} and 22 mm d^{-1}, respectively. Variations in measured climatic variables were found to account for ~90% of variation in ablation during both summer and winter seasons, with significant increases in ablation occurring in conjunction with heavy rainfall events. Surface velocity also showed seasonality, averaging 0.87 m d^{-1} during summer and 0.64 m d^{-1} in winter, a reduction of ~26%. It is thought that the general reduction in velocity during winter can be attributed to a decrease in the supply of surface meltwater to the subglacial zone. Short-term velocity peaks appeared to coincide with heavy rainfall events, with surface velocity responses typically occurring within 24 hours of each rainfall event. During winter, moderate rainfall events (≤100 mm over 24 hours) created a surface velocity response up to 44% greater than the prevailing velocity. Though difficult to deconvolve, magnitudes of surface velocity response to rainfall inputs appear linked to time lags between rainfall events and subglacial drainage efficiency and water storage. The longer-term dynamics of Fox Glacier appear linked to fluctuations in the Southern Oscillation Index (SOI), with positive mass balances of Southern Alps' glaciers appearing to mirror negative SOI (El Niño) conditions. Given the calculated response time of ~9.1 years for Fox Glacier, the current terminus advance may be linked to mass gains reported in the mid-1990s, with current mass balance gains perhaps leading to terminus advances ~9 years hence.

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Introduction

Intra-annual variation in ablation has often been identified on valley glaciers, and this is controlled by net radiation and sensible heat contributing energy for surface melt (Ishikawa et al., 1992; Marcus et al., 1985; Owens et al., 1992). Seasonal fluctuations in surface velocity have been identified on glaciers in Europe and North America (Anderson, 2003; Hooke et al., 1989; Paterson, 1964; Willis, 1995) but not as yet on New Zealand glaciers. Surface motion is often linked to variations in water supply to the glacial drainage system, where increases in water pressure in subglacial channels and conduits result in increased basal sliding as the ice/bed interface decouples (Hooke et al., 1989; Mair et al., 2001; Nienow et al., 1998; Willis et al., 2003). In particular, short-term variations in surface velocity have been interpreted in terms of variations in basal motion, itself influenced by augmented subglacial water storage (Iken et al., 1983; Jansson, 1995) and basal water pressure (Kamb et al., 1994; Meier et al., 1994). New Zealand's maritime climate results in warm-based glaciers at the pressure melting point, and large quantities of meltwater, allowing basal sliding to occur throughout the year (Ruddell, 1995). In addition, due to the low altitude of glaciers on the west coast (termini are around 300 m above mean sea level), heavy rainfall events can occur throughout the year. Rainfall events have been found to significantly increase ablation as latent heat is released from the freezing of raindrops on contact with ice (Ishikawa et al., 1992; Marcus et al., 1985; Takeuchi et al., 1999). Such events have been related to short-term variations in surface velocity (Andreasen, 1983; Goodsell, 2005).

Fox Glacier (Fig. 1) has been the focus of little scientific research, with the majority of research being conducted on the neighboring Franz Josef Glacier (which terminates 12 km to the northeast). The premise has been that that the two glaciers, due to their close proximity, size, and shape, exhibit similar behavior (Sara, 1968), with extensive climate records from Franz Josef township allowing detailed temporal analyses of climate-glacier interactions on Franz Josef Glacier (Hooker and Fitzharris, 1999). Prior to the present study only one published study of ablation had been conducted on Fox Glacier (Gunn, 1964), with no study of winter ablation rates. More attention has been given to surface velocity (Gunn, 1964; Speight, 1935; Wilson, 1896), but measurements have not been conducted for 13 years, and no comparison of summer and winter velocities has ever been made. Fox Glacier has undergone rapid retreat since the Little Ice Age and particularly during the first three-quarters of the 20th century (Chinn, 1996). Since Sara's (1968) survey, remote sensing using aerial photography has been the only method used to delineate the terminus position of Fox Glacier, and this indicated a large advance of the terminus through the 1980s and 1990s (Chinn, 1996), culminating around 1999 (Coates and Chinn, 1999). Since then, the terminus position has appeared to be relatively stable.

Valley glaciers are considered to be sensitive indicators of medium- to long-term climatic change, although individual glacier response can be extremely complicated due to variables including

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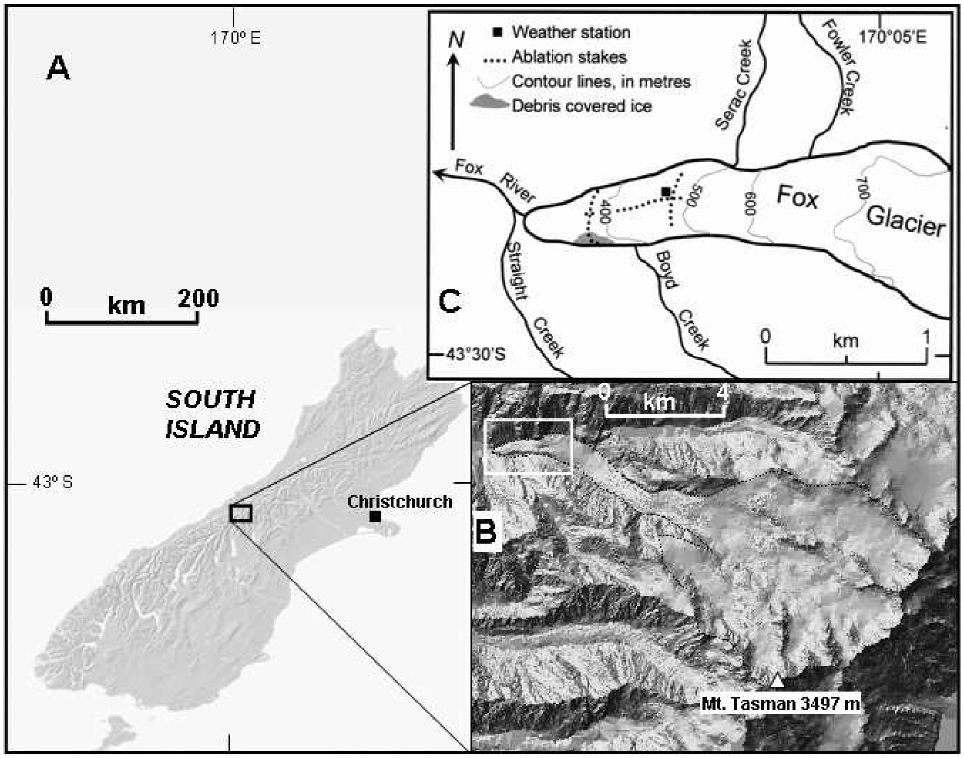

FIGURE 1. (A) Location map of Fox Glacier on the South Island. (B) Shaded relief map of Fox Glacier (from Land Information New Zealand 20 m contour data); white rectangle refers to area in C. (C) Location of climate station and ablation stake transects used for both ablation and surface velocity measurements.

the nature of climate forcing, valley geometry, basal hydrology, and the mechanics of glacier flow (Hubbard, 1997). Response time is broadly defined as the time taken for a glacier to adjust over its entire length and volume to a change in mass balance. Although response time is one of the most important physical variables characterizing glacier dynamics, it has been difficult to define analytically, and has consequently been the focus of much research (Lliboutry, 1971; Nye, 1963; Paterson, 1994). Jóhannesson et al. (1989) found a strong linkage between response time and the volume time-scale changes of a glacier (Hubbard, 1997), with their results indicating that response time calculated with Nye's (1963) kinematic wave theory unnecessarily complicated. Response time for valley glaciers is now thought to be of the order of 10^1–10^2 years (e.g. Jóhannesson et al., 1989; Oerlemans, 2001), substantially less than estimated from earlier methods (e.g. Paterson, 1981). Response times have previously been calculated for Franz Josef Glacier but not for Fox Glacier and these results range from 5 to 25 years depending on the technique adopted (Evans, 2003; Jóhannesson et al., 1989; Oerlemans, 1997; Paterson, 1994). Coates and Chinn (1999) believe that Fox Glacier's response time lags that of Franz Josef Glacier by one year and have postulated a response time of 6 years for Fox Glacier. If this one-year lag exists, and considering the varied response times reported above, a response time for Fox Glacier should be of the order of 6–26 years. Given the lack of glacier-climate research over seasonal and decadal time scales at Fox Glacier, the aims of this study are twofold: (1) to determine factors driving seasonal variations in ablation and velocity at Fox Glacier; and (2) to explore linkages between glacier response time and El Niño–Southern Oscillation. Findings will also be compared with results from published studies from the neighboring Franz Josef Glacier.

Study Area

Fox Glacier is a temperate valley glacier located on the South Island's west coast at 43°30′S and 170°10′E (Fig. 1). Fox Glacier's névé at 2700 m a.s.l. is one of the largest in New Zealand in terms of area, encompassing a collection area of 25 km^2 (Anderson, 2003). From this large catchment the glacier descends steeply over a distance of 12.7 km, terminating at 270 m a.s.l. only 17 km from the present coastline. The glacier's catchment area intercepts the dominant westerly airflow, resulting in up to 15 m of rainfall per year near the main divide of the Southern Alps (Coates and Chinn, 1999). October is the wettest month and June to August are the driest months, though no published information exists as to variations in solid precipitation (snow) and rain between the two seasons (NIWA, 2005).

Methods

Ablation was measured via a stake network with twenty-five 2-m-long, light gray PVC tubes placed into hand-augered holes in the ice. The stake network comprised two 200-m-wide transverse profiles intersecting a longitudinal profile, with stakes approximately 30–40 m apart (Fig. 1C). Unfortunately, due to heavy crevassing and rock fall activity at the margins of the glacier during both the summer and winter field seasons, the stake network extended only to within ~70–80 m of the margins. Hence, this precluded an extensive analysis of surface velocity pattern, such as transitions between normal (parabolic) and plug-like surface flow following rain inputs. Ablation at each stake was read daily at the same time, by the same operator, by measuring the height of the exposed stakes at the ice surface, using a tape to the nearest millimeter (Hubbard and Glasser, 2005). Microtopography may develop around stakes due to the conductivity of the PVC, making ablation measurements potentially erroneous. To address this, the standard straight edge technique (outlined in Müller and Keeler, 1969) was used, which has a measurement error of ±5 mm. Surface velocity measurements utilized the same stake network (Fig. 1C), which was set up with one central longitudinal transect and two transverse transects. Stake positions were recorded daily with a Trimble R8 real-time-kinematic differential global positioning system (RTK-dGPS; e.g. Hubbard and Glasser, 2005), giving horizontal (*x, y*) precision of ±1 mm and vertical (*z*) precision of ±20 mm. The Rover unit was fitted directly into the top of each stake using a bracket, and stakes were held steady and vertical during measurement, diminishing the likelihood of measurement errors, following Anderson (2003). Satellite coverage was found to be most comprehensive between 10:00 and 12:00 in the Fox Glacier area during both field seasons, so RTK-dGPS measurements were always made during this time period. Because ice moves upward toward the surface in the ablation zone, net vertical change can be less than estimates made from the movement of surface features. This is called the *emergence velocity*, and is the rate that the ice surface would rise at if there were no ablation (Paterson, 1994). Hence, velocity was calculated by incorporating both changes in horizontal and vertical components, and also ablation at each survey point (e.g. Paterson, 1994). To calculate velocity from the GPS method, the change in the eastings, northings, and altitude (including change due to ablation) over the survey time period (i.e. hours, days, weeks) can be calculated (Hubbard and Glasser, 2005). Measurements were conducted for two periods in 2005, from 9 January to 3 February and from 8 June to 3 July. During each study period a weather station with a Campbell Scientific CRX10 data logger elevated 1.5 m above the ice surface in a position central to the stake network recorded temperature, humidity, precipitation, wind speed, wind direction, and incoming short-wave solar radiation. The climate variables were recorded at 15 minute intervals from which daily mean values were calculated. The results were then related to ablation and surface velocity measurements.

Results

ABLATION

During summer, ablation on the clean ice surfaces ranged between 50 and 216 mm d^{-1}, and averaged 129 ± 81.1 mm d^{-1} (±2 standard deviations), with higher ablation rates on the lower part of the longitudinal transect. In winter, the ablation ranged from –2 (solid accumulation day) to 83 mm d^{-1} with an average of 22 ± 23.4 mm d^{-1} (±2 standard deviations), or only 5 mm d^{-1} if days with rainfall are excluded (13 days; Fig. 2). Using stepwise multiple regression, it was found that variations in measured climate parameters accounted for 89.9% of the ablation variability (*A*) recorded during both summer and winter monitoring (Purdie, 2005):

$$A = 6 + 9.17_{\text{temp.}} + 0.448_{\text{precip.}} - 0.385_{\text{humid.}} + 2.19_{\text{windsp.}} + 0.151_{\text{solarrad.}} \quad (1)$$

Large spatial variability was recorded between the ablation rates on the clean ice surfaces and that occurring under the debris covered ice on the lower true left of the glacier (Fig. 1). This was particularly significant during summer when the ablation rate under the debris-covered area surveyed was suppressed by up to 50%. The debris cover in this zone was ≥10 mm thick, allowing insulation of the ice surface beneath. Indeed, summer ablation values ranged from 3 to 143 mm d^{-1}, with a daily mean of 64 mm d^{-1} (±32.7 mm d^{-1}). In winter, ablation on debris-covered ice was again generally lower than on clean ice, with a range of 1 to 57 mm d^{-1} and a mean daily value of 13 mm d^{-1} (±13.0 mm d^{-1}). The large standard deviations are manifestations of the differing debris thicknesses. Spatial variability may also be related, in part, to surface topography. Indeed, this may be the case for one stake sited on the lower transect to the immediate true left of the longitudinal transect. This stake was located on the western slope of a surface depression, with a southerly aspect (in contrast to a westerly or northerly aspect for the rest of the stake network), and had the lowest daily ablation rate of any measured area during summer field season (a daily average ablation rate of 8.45 mm d^{-1}). This contrast due to topoclimatic effects was not identified during winter due to the lack of direct insolation onto the lower glacier presumably as a result of topographic shading.

VELOCITY

During the summer the overall average daily surface velocity across the stake network was 0.87 ± 0.27 m d^{-1} (± 2 standard deviations), and in winter only 0.64 ± 0.24 m d^{-1}, equating to a 26 ± 7% reduction in surface velocity during the winter season (Fig. 2B). Coupled with reduced ablation, this winter decrease in glacier motion led to a 3 m thickening of the glacier snout, due to emergence velocity, the upward motion of ice (c.f. Paterson, 1994). Although the final day of the winter field season yielded particularly high velocity results of up to 1.30 m d^{-1} (mean of 1.07 m d^{-1}), this actually has a minimal effect on the overall results for the winter field season. Indeed, if the final day's velocity results are excluded from the analysis, overall winter velocities decrease only slightly to 0.61 ± 0.17 m d^{-1}. The limits of the cross-glacier transects make it difficult to draw unequivocal conclusions as to the nature of the relationship between hydrological processes and velocity. However, the limited seasonally averaged velocity fields are suggestive of a parabolic pattern of surface velocity, with a broad area of relatively high velocities close to the center line and a decrease in velocity near the margins (Purdie, 2005). Short-term variations in surface velocity do appear to be linked to variations in water supply, with heavy rainfall events (≥100 mm) leading to velocity increases ranging from 4 to 44% over intervals of 1–2 days (Fig. 2). The most synchronous response between increased water input and surface motion response occurred on 22 January. On this occasion, increased temperatures (14°C) coupled with a large rainfall event (120 mm over 24 hours) led to a substantial increase in surface motion (Fig. 2A). The smallest percentage increase in surface motion occurred in winter, following a rainfall event on 19 June (Fig. 2). This rainfall event itself followed one of similar magnitude only 3 days prior.

RESPONSE TIME OF FOX GLACIER

Jóhannesson et al. (1989) developed two simple models to estimate the general response time of a glacier, one utilizing glacier length and terminus velocity (Equation 2), the other utilizing glacier thickness and annual ablation at the terminus (Equation 3). The first technique is:

$$T_M = fl/u, \quad (2)$$

where T_M is the response time, *f* a factor estimated to be 0.5, *l* the glacier length (12.7 km) and *u* the terminus velocity (the average of summer and winter mean velocities is 0.72 m d^{-1} or 263 m a^{-1}). The second technique of Jóhannesson et al. (1989) is:

$$T_M = H/-b, \quad (3)$$

where *H* is glacier thickness and –*b* annual ablation at the terminus. The precise thickness of Fox Glacier is unknown, and

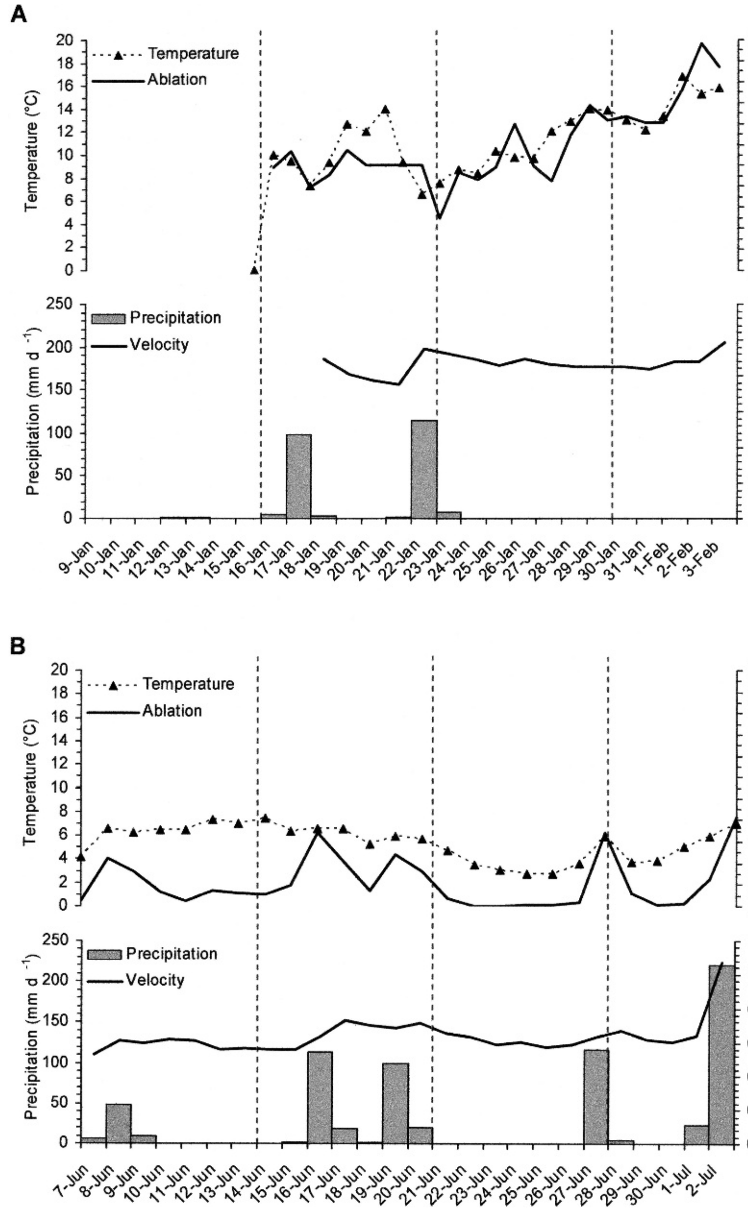

FIGURE 2. Intra-annual variations in average daily ablation (mm d^{-1}), precipitation rate, daily ice flow velocity, and temperature; (A) summer, (B) winter.

although radio-echo sounding has been used elsewhere to determine ice thickness more accurately, Ruddell's (1995) thickness estimate of 200 m is based on the Fox Glacier surface elevation and the topography of the surrounding valley. Utilizing the data gathered during this study and an estimated average thickness of 200 m (Ruddell, 1995), the response time of the Fox Glacier using Equation 2 is 24.1 years, and with Equation 3, 9.1 years. Nevertheless, both techniques are crude methods of both temporal and spatial response of a glacier to mass balance perturbations, and a lack of precise thickness variations can make conclusions potentially spurious. For the first method (Equation 2), f is the shape factor used in the calculation of basal shear stress and is related to valley shape. It can only really be inferred to be 0.5 if the half-width of the glacier is equal to the depth (the depth of Fox Glacier has only been estimated). Notwithstanding this limitation, based on the 200 m depth estimate of Ruddell (1995), a shape factor of 0.5 would appear to be justified, given that the half-width of the lower 2 km of Fox Glacier is ~200 m (Fig. 1C).

Discussion

ABLATION

Seasonal variations in both ablation and surface velocity were detected. Indeed, during summer, ablation on clean ice surfaces ranged between 50 and 216 mm d^{-1}, and averaged 129 ± 81.1 mm d^{-1}. In contrast, in winter, ablation ranged from −2 (a solid accumulation day) to 83 mm d^{-1} with a mean of 22 ± 23.4 mm d^{-1} (only 5 mm d^{-1} if days with precipitation are excluded). During summer, average daily surface velocity was 0.87 ± 0.27 m d^{-1}, while in winter this figure decreased to 0.64 ± 0.24 m d^{-1}, equating to a 26 ± 7% reduction in surface velocity during the winter season. During summer, temperature, humidity, and solar radiation were the most important parameters influencing ablation, a reflection of the importance that both radiation and sensible heat fluxes have on the energy available for melt (Braithwaite, 1981), while in winter a strong positive relationship was identified between ablation and rainfall. This demonstrates how latent heat released from the freezing of raindrops on contact with ice, although usually considered to be less important, can play a significant role in maritime climates (Marcus et al., 1985; Takeuchi et al., 1999). Clean ice ablation rates indicate that annual ablation is in the vicinity of 22 m water equivalent (w.e.), a similar rate to the 20 m w.e. recorded on nearby Franz Josef Glacier (Anderson, 2003). These figures are much higher than ablation rates recorded on maritime Norwegian glaciers (10 m w.e.; Oerlemans, 2001), and ablation rates on glaciers in continental climates where annual ablation rates are only around 2 to 3 m w.e. (Oerlemans, 2001). Average summer ablation rate (129 mm d^{-1}) is within the range of summer ablation rates reported for Franz Josef Glacier. Indeed, Owens et al. (1992) reported a figure of 137 mm d^{-1}, while Gunn's (1964) rate (82 mm d^{-1}) and Evans' (2003) rate (77 mm d^{-1}) are slightly lower. In winter, our average rate of 22 mm d^{-1} appears slightly larger than the winter ablation rate of 12 mm d^{-1} reported for Franz Josef Glacier by Marcus et al. (1985). However, if days with rainfall are excluded from the analysis, our winter ablation rate at Fox Glacier decreases to 5 mm d^{-1}. This is much less than the winter ablation rate reported for Franz Josef Glacier, and potentially could be due to the more westerly aspect of Fox Glacier (compared to northwesterly for Franz Josef Glacier). As a result, during winter the lower part of Fox Glacier is topographically shaded for most of the day, allowing less surface melt.

Consistent with previous research, the amount of ablation suppression increased as debris thickness increased (Nakawo and Young, 1981). Indeed, at some stakes, the presence of a layer of smaller clasts (≤5 mm *a*-axis) at the ice interface, which were overlain by larger clasts, appeared to enhance the insulation properties of the debris cover. Pelto (2000) has also observed how a finer-grained debris cover at the ice interface, overlain by a layer of larger clasts, may have a greater insulating capacity. However the precise dynamics of this effect are unknown, and although variations in clast size at each of the five stakes in debris-covered ice were noted, the actual thickness of this fine layer was not quantified. Clearly, further research into the influence of this layer on the total insulation capacity of the entire debris layer would be intriguing, but is beyond the scope of this paper.

During the winter, debris cover was also found to have a sheltering effect on the underlying ice during the four heavy rain events (≥100 mm), with rates of ablation being suppressed by about 44% in comparison to that occurring on the clean ice surface. This reduction in ablation is likely to be related to decreased raindrop impact due to the protecting debris cover, diminishing possible thermal effects of rain interacting with the ice surface (e.g. Paterson, 1994).

VELOCITY

Seasonal velocity ranges are similar to those recorded on glaciers in the northern hemisphere, where velocity variations of up to 30% have been recorded. These have been related to changes in basal water supply via rain events and surface melt (e.g. Hooke et al., 1989; Kamb and Engelhardt, 1987; Willis et al., 2003), although Kamb and Engelhardt's (1987) study was on the surging Variegated Glacier, which does not show typical seasonal variations. During the winter season on Fox Glacier, surface motion deceleration can also be attributed to a decrease in water supply, resulting in lower basal sliding rates. During winter, the lower Fox Glacier is topographically shaded, average daily temperature is only 5.4°C (c.f. 11.5°C in summer), and average daily ablation is reduced by 83%, contributing to a reduced water input. The magnitude of velocities recorded during this study (250 m yr^{-1}) indicates that Fox Glacier has similar surface flow rates to other maritime glaciers of this size (Benn and Evans, 1998).

There appeared to be variations in the magnitude of velocity response to water inputs (rainfall), and also slight variations in the time taken for a response to occur, ranging from instantaneous surface motion responses, to at least a 24 hour delay (Fig. 2). Surface motion responses to moderate increases in water input in the winter survey period (Fig. 2B) were detectable, though more minimal than in summer (Fig. 2A). This could reflect the fact that the rain events on 16, 19, and 27 June were moderate in magnitude (90–100 mm over 24 hours), compared with the event on 2 July (220 mm over 24 hours), which itself produced a substantial surface motion response. The corollary may be that even during winter, the subglacial drainage network appears reasonably well evolved enough to evacuate increased water inputs of <100 mm falling over a 24 hour period, without a major surface velocity response occurring. Elapsed time between discrete rain events (≥3 days) may also be an influence on the magnitude of the surface velocity response. In winter, on the three occasions where a significant surface velocity response occurred (16 and 27 June and 2 July; Fig. 2B), there was a short time lag (<2 days) between increased water input and velocity response. On these occasions, there had not been any rain for at least 5 consecutive days. In winter, this slight delay in surface velocity responses to water inputs may be due to the lack of the contributing effect of surface melt, or to storage levels in cavities being low enough to allow moderate increases in water input before bed separation occurs (Iken et al., 1983). In summer, surface melt coupled with a very large rain event led to an instantaneous velocity response (22 January). This is in spite of a presumably highly efficient drainage network at that stage of the melt season, as increasingly efficient, channelized drainage systems often develop within and beneath alpine glaciers as the melt season progresses (e.g. Nienow et al., 1998). In addition to rainfall driving velocity increases, a minor surface velocity response (11%) was recorded around 3 February (Fig. 2A). This coincided with a temperature increase of around 3°C and an ablation increase of ~70 mm d^{-1}, suggesting a direct relationship between ablation (in particular surface melt followed by runoff) and velocity. The generally short time lags (<2 days) between changes in water input and surface motion response concurs with findings from studies of alpine glaciers in the northern hemisphere. Indeed, Mair et al. (2001, 2003) found very short time lags between increased water inputs and surface motion responses, especially during melt season "spring events."

The average summer velocity that we report (0.87 m d^{-1}) is similar to the average summer velocity of 0.70 m d^{-1} for Fox Glacier reported by Ruddell (1995), within the range of results for nearby Franz Josef Glacier (spatially variable rates of 0.2 to 1.0 m d^{-1}) reported by Anderson (2003), and at the lower end of the range (spatially variable rates of 0.73 to 1.48 m d^{-1}) reported by McSaveney and Gage (1968). In contrast to our study of Fox Glacier, seasonal variations in surface velocity have not been detected at Franz Josef Glacier (Anderson, 2003). This lack of seasonality has been linked to the northwesterly aspect of the snout of Franz Josef Glacier (Goodsell et al., 2005), which is in contrast to the westerly aspect of the tongue of Fox Glacier. As a result, there may be less seasonal contrast in the supply of meltwater to the glacier bed (due to surface melt) at Franz Josef Glacier than at Fox Glacier.

RESPONSE TIME

Both methods of calculating response time (T_M) that we implemented have their limitations and make interpretations and

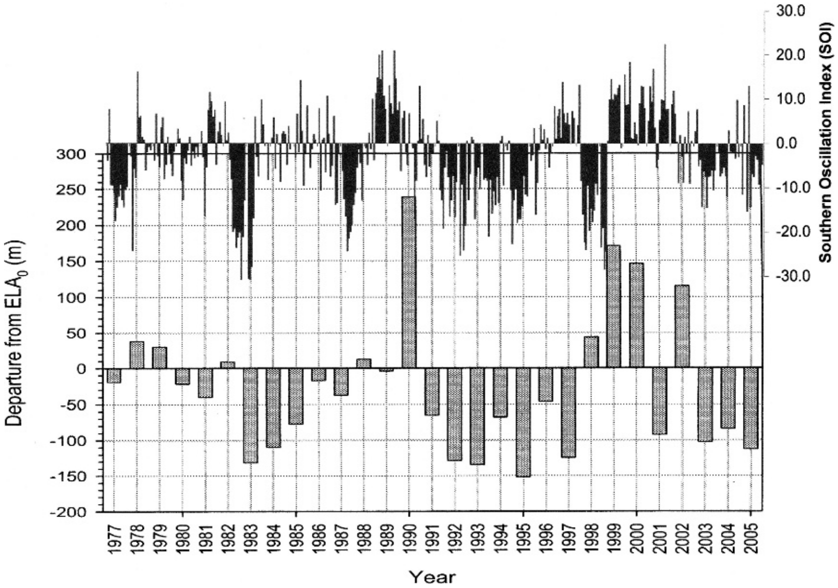

FIGURE 3. Mean annual departures (bottom) from the steady state ELA (equilibrium line altitude) as monitored by NIWA (National Institute of Water and Atmospheric Research) in the annual snowline survey (Chinn et al., 2005a), overlain by variations in the Southern Oscillation Index (top) as recorded by the Australian Government Bureau of Meteorology (Bureau of Meteorology, 2005). Patterns of El Niño (negative SOI) years show similarity to those with negative snowline departures that correspond to positive mass balance.

conclusions only tentative (Hubbard, 1997). Indeed, the methods are based on a single, specified mass balance perturbation, constant in time, together with an estimation of glacier thickness based on the elevation of the glacier surface compared with the surrounding topography. A positive mass balance feedback, whereby the net mass balance across a glacier tends to increase as surface elevation increases, makes this assumption unrealistic and will tend to extend response time. Response time results using the two methods mirror the findings of Jóhannesson et al. (1989), specifically that Equation 2 exaggerates response time. Indeed, using Equation 2, response times have previously (e.g. Paterson, 1994) been found to be of the order of 10^2–10^3 years (for Fox Glacier we calculated 24.1 years). This is the theoretical, long time scale commonly thought to be representative of typical glaciers, but Jóhannesson et al. (1989) cautioned that using this method, response times appear longer than available observations would indicate. Consistent with Jóhannesson et al.'s (1989) second model (Equation 3), a much lower response time of 9.1 years was calculated, and consistent with Jóhannesson et al. (1989), this much lower value calls into question the usefulness of the theoretical long time scale method (Equation 2). Nevertheless, preferring one response-time technique over another is fraught with difficulties, especially given the lack of reliable glacier thickness data.

There is growing evidence for teleconnections between glacier response and atmospheric circulation patterns in both the northern and southern hemispheres (Hooker and Fitzharris, 1999). Indeed, maritime glaciers in western Norway and the western side of the Southern Alps (e.g. Fox Glacier) have both experienced recent glacial advances (Chinn et al., 2005b). Common to both countries, these positive mass balances are associated with the strength of westerly airflow, which brings increased precipitation. Previous research has correlated glacier advance in New Zealand with changes in the Southern Oscillation Index (SOI), in particular mass gain during periods of a negative SOI associated with El Niño conditions (Clare et al., 2002; Hooker and Fitzharris, 1999; Tyson et al., 1997). In the South Island of New Zealand, El Niño seasons are characterized by generally strengthened westerly circulation in summer, southerly circulation in winter, and southwesterly circulation in autumn and spring, with lower than normal temperatures (Chinn et al., 2005b). Fluctuations in the SOI (Fig. 3) can be seen to have remarkable similarity to the recorded mean annual snowline departures for Southern Alps glaciers. Indeed, the plot of SOI and departures from ELA (equilibrium line altitude) (Fig. 3) would appear to confirm a correspondence between the general negative departures from ELA (mass gain) and negative SOI (El Niño) conditions. During mass gains, anomalous high pressure exists to the south of Australia, with lower than average air pressure over New Zealand. Stronger westerlies create higher than average precipitation, and melt is restricted by cooler temperatures (Hooker and Fitzharris, 1999), promoting positive glacier mass balances and terminus advances. The ~200 m advance of Fox Glacier since the mid-1980s (Chinn et al., 2005a) and the positive mass balance state of neighboring Franz Josef Glacier (Anderson, 2003) indicate Fox Glacier is currently in a state of positive mass balance. This appears to relate to negative (El Niño) SOI conditions (Fig. 3), and the two response times (T_M) calculated for Fox Glacier can be compared with the SOI record. The 9.1 year response time estimate compares favorably with data gathered by the National Institute of Water and Atmospheric Research (NIWA) in their annual glacier end-of-summer snowline survey (EOSS; Chinn et al., 2005a). Figure 3 shows that in 1995 there was an average depression of the EOSS by 150 m in comparison to the steady state ELA, and from 1992 through to 1997 a general increase in mass over the glaciers monitored along the Southern Alps, in concert with several negative SOI years. The 24.1 year response time estimate does not appear to closely mirror the SOI/EOSS data in Figure 3. The recent mass gains of 2003–2005 should be reflected in a series of negative SOI years in the late 1970s to early 1980s, and depression of EOSS. However, only a very weak negative SOI prevailed from 1978 to 1982, and for 3 of those years EOSS were elevated (Fig. 3). Accepting the limitations of the response time methods, if the response time of Fox Glacier is ~9 years, the mass gain from El Niño events of the mid-1990s may be linked to the current (2005) terminus advance, with the mass gains in the mid-1980s responsible for the 1990s advance that culminated in 1998/1999 (Chinn et al., 2005b). The corollary is that the full impact of recent mass gains (2003–2005) may not begin to be expressed at the terminus until ~2012.

Conclusions

Large seasonal variation exists in ablation on the lower Fox Glacier, with the average winter ablation rate (22 mm d^{-1}) being roughly one-sixth of the average summer ablation rate (129 mm d^{-1}). However, it appears significant ablation can occur at any time of the year in association with heavy rainfall events. Debris cover was found to significantly suppress ablation (50%) and also provided a sheltering effect from heavy rainfall events. The winter surface velocity was considerably (26%) lower than in summer, which agrees with previous studies of valley glaciers elsewhere (e.g. Hooke et al., 1989; Mair et al., 2001; Nienow et al., 1998). Topographic shading due to the constraining valley walls and cooler temperatures during winter appear to control surface melt contributions to basal sliding. Although ablation rates and velocity are similar in magnitude to results reported for Franz Josef Glacier, the seasonality of velocity at Fox Glacier would appear to contrast with the neighboring Franz Josef Glacier (Anderson, 2003). It is uncertain why this is so, although it could be due to the differing aspects of the tongues of the two glaciers, with the northwest-facing Franz Josef Glacier receiving longer durations of direct sunlight in winter. Short-term velocity peaks appeared to coincide with heavy rainfall events, with surface velocity responses typically occurring within 24 hours of each rainfall event. The magnitude of, and time lag of, velocity response to rainfall inputs may be linked to drainage efficiency and water storage at the glacier bed, which control basal sliding. Positive mass balance gain of glaciers in the Southern Alps appears to mirror negative SOI (El Niño) conditions, and given the calculated response time of ~9.1 years for Fox Glacier, the current terminus advance of Fox Glacier could be linked to mass gains reported in the mid-1990s. The implication is that current mass balance gains (Chinn et al., 2005a) may well lead to terminus advances ~9 years hence.

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