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Response of Franz Josef Glacier *Ka Roimata o Hine Hukatere* to climate change

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ABSTRACT

Understanding the response of relatively small ice masses to climate change on a century-to-century timescale is important because of their significant contribution to sea level change in the recent past, and their likely contribution in the next century. Franz Josef Glacier, with its excellent terminus position record, location in the Southern Hemisphere, and its extreme sensitivity to climatic changes, is a key data point for understanding the magnitude and mechanisms of past and future climate changes. Here, we use a reconstructed mass balance record to drive an ice flow model which simulates the general retreat of the glacier during the 20th century. The model is then used to project future glacier variations, based on regionally downscaled climate change scenarios. The results indicate that, under a mean climate change scenario, the glacier will retreat 5 km and lose 38% of its mass by 2100. Different scenarios give a range of recession from 3.9 to 6.4 km, and a range of volume loss from 26% to 58%. On a global scale, the magnitude of the retreat is at the upper end of the range calculated for other, similar sized, glaciers.

1. Introduction

Understanding the response of relatively small ice masses to climate change on a century-to-century timescale is important because of their significant contribution to sea level change in the recent past, and their likely contribution in the next century. Changes in the volume of small ice masses accounted for 25% of the measured rise in eustatic sea level between 1993 and 2003 (IPCC, 2007), and is likely to account for 33% of sea level rise by 2100 (Gregory and Oerlemans, 1998; van de Wal and Wild, 2001; Houghton et al., 2001). The high sensitivity of small ice masses to climate change is a result of their small system scale, and their relative proximity – compared to the major ice sheets – to the melting point. At a local scale, small ice masses are often important community resources: water from glacierised basins is used for hydro-electric generation and irrigation, and in some areas glaciers are key drivers of local tourism industries (Braun et al., 2000).

Within that context – of relatively high sensitivity of small ice masses – the Franz Josef Glacier is extremely sensitive. It has ablation rates of up to 20 m a^{-1} w.e., accumulation rates of up to 8 m a^{-1} w.e. (Anderson et al., 2006), and its terminus position fluctuates rapidly. It also has the best record of terminus position of any glacier in the Southern Hemisphere (Grove, 2004). It is therefore a key locale for developing an improved understanding of century scale climate change response processes in small glaciers.

As one of the best studied Southern Hemisphere glaciers, Franz Josef Glacier is a key indicator of Southern Hemisphere climate on a number of timescales. On a geological timescale, the timing and magnitude of glacial advance helps understanding of the mechanisms of global climatic changes through glacial cycles. More recent advances, such as during the Little Ice Age, provide information on millennial-scale climatic variations (McKinzey et al., 2004). During the 20th century glacial retreat indicates a general warming trend (Oerlemans, 2001) with smaller readvances caused by cooler, wetter conditions linked to the El Niño Southern Oscillation and Inter-decadal Pacific Oscillation (Hooker and Fitzharris, 1999; Chinn et al., 2005).

The objective of this study is to examine the response of the Franz Josef Glacier to climate change in the past and in the future. Specific goals are to:

1. test the ability of a coupled mass balance – glacier flow model to simulate glacier advance and retreat for the period for which instrumental climate data exist, 1894–2005.
2. generate scenarios of long term glacier behaviour to 2100 using future climate scenarios constructed within the IPCC framework and downscaled to the New Zealand environment.
3. compare the response of Franz Josef Glacier to that of glaciers in different environments.

These objectives will be realised by the application of a glacier flow model to the Franz Josef Glacier. This model will be verified by comparison with the record of terminus position and measurement of ice velocity. Ice thickness, and temporal and spatial variations in ice velocity have been measured by Anderson (2004) and provide input and tuning data for the model. The mass balance input to the model is

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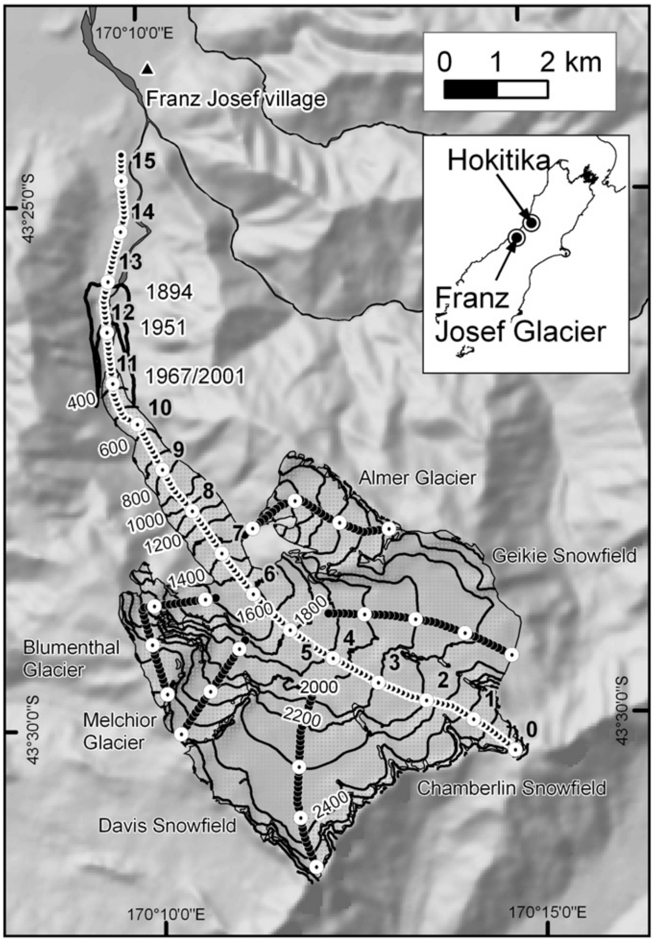

Fig. 1. Map of Franz Josef Glacier and its location on the West Coast of the South Island of New Zealand. Selected terminus positions from 1894 to 2001 are mapped. Contours are in m above sea level and the different parts of the glacier which contribute to the glacier tongue are labelled. The flowlines used to parameterise the glacier geometry are shown, with a grid spacing of 100 m, and numbers indicating kilometres from the glacier head. The main flowline is not glacierised past 11.5 km (2001), and the Almer flowline after 3 km.

derived from a degree–day mass balance model constructed by Anderson et al. (2006) which has been evaluated against measured mass balance.

2. Franz Josef Glacier

The Franz Josef Glacier/*Ka Roimata O Hine Hukatere* is ~11 km long and occupies 35 km^2 on the western flanks of the Southern Alps/*Ka Tiritiri O Te Moana* of New Zealand/*Aotearoa* (Fig. 1). The maximum elevation of the glacier is 2900 m a.s.l., although the area at this elevation is small, consisting of snow and ice perched on steep mountain sides. Most of the accumulation area consists of broad gently sloping snowfields, from 2600 m a.s.l. to 2000 m a.s.l. Between 2000 and 1700 m a.s.l., ice from these snowfields converges in a heavily crevassed area. Below 1700 m a.s.l., the glacier tongue is steep and narrow, consisting of a series of icefalls.

Precipitation on the glacier increases from c. 7 m a^{-1} at the terminus to c. 12 m a^{-1} at 600 m a.s.l. before reducing to c. 5 m a^{-1} at the head of the glacier, while mean annual temperature varies from c. 9 °C at the terminus to c. −4 °C at the head of the glacier (Anderson et al., 2006). The resulting high accumulation and ablation rates in turn lead to ice velocities in the order of 1000 m a^{-1} (Anderson, 2004), and response times estimated between 9 and 20 yr (Oerlemans, 1997a). The highly dynamic nature of this system is most obvious when looking at the terminus position record of the glacier.

Between the first survey of the glacier in 1894 and the 20th century minimum in glacier length, recorded in 1984, the glacier retreated c. 3 km as a result of a regional warming estimated from climatic records of c. 1 °C (Salinger et al., 1995). Within this general retreat, there have been a number of smaller readvances, the latest of which started in 1984 and is continuing today, despite a retreat of ~450 m from 1999–2005.

3. Glacier flow model

The ice flow model used in this study is a simple one-dimensional flowline model solved on a finite difference grid. Variations of this model have been used by many authors (e.g. Oerlemans, 1988; Greuell, 1992) and the model will be described only briefly here. The model is an improvement over those used previously on Franz Josef Glacier, by the use measured ice thickness, multiple flowlines and the use of a 100 m grid spacing. Recent velocity measurements by Anderson (2004) provide a sound basis for flow parameter choice.

The required inputs for the model are an initial glacier geometry, and mass balance to force the model. The outputs are ice thickness, terminus position and ice velocity.

3.1. Description

The dynamic behaviour of the glacier is described in terms of change in ice thickness which is calculated from a continuity equation (Paterson, 1994). Glacier velocity is calculated from basal shear stress with contributions from internal deformation and basal sliding.

The glacier cross cross-section S is represented by a trapezium, described by the ice thickness, H the width of the bed w_b, and the slope of the valley sides θ, here taken as being equal on both sides (Fig. 2).

The width at the surface w_s is given by:

$$w_s = w_b + 2H\tan\theta \tag{1}$$

and the area of the trapezium S by:

$$S = H(w_b + H\tan\theta). \tag{2}$$

Assuming constant density, the continuity equation is integrated over the transverse cross-section to give the change in cross cross-section area due to mass balance and ice transport (Paterson, 1994):

$$\frac{\partial S}{\partial t} = \frac{\partial}{\partial x}(US) + w_s B \tag{3}$$

where B is the specific mass balance, t is time and U mean velocity through the cross cross-section. The equation is reformulated in terms of ice thickness change by substituting Eq. (2) into Eq. (3) (Budd et al., 1979; Paterson, 1994):

$$\frac{\partial H}{\partial t} = \frac{1}{w_b + 2H\tan\theta}\frac{\partial}{\partial x}(US) + B \tag{4}$$

where U and S both depend on H. The mean cross-section velocity is calculated from the local driving stress approximation (Paterson, 1994):

$$U = U_d + U_s = f_d H \tau_{xz}^n + f_s \frac{\tau_{xz}^r}{H} \tag{5}$$

where U_d is the velocity due to deformation and U_s the velocity due to basal sliding. The constants f_d and f_s are the deformation and sliding

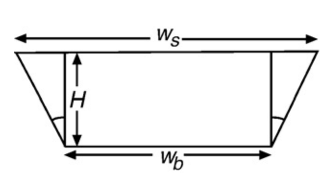

Fig. 2. Glacier cross cross-section representation.

parameters respectively. The exponent in Glen's flow law n, is taken as 3 for this study. The sliding law exponent r is also taken as 3, a number which arises from the analysis of viscous ice flow around bumps. Changes in basal water pressure are a major cause of temporal variations in sliding velocity, but are neglected here because of the difficulties in water pressure calculation. The bulk effect of basal water pressure is incorporated through the sliding flow parameter.

The basal shear stress τ_{xz} is calculated from the local driving stress only:

$$\tau_{xz} = -F\rho gH\sin\alpha \tag{6}$$

where ρ=910 kg m^{-3} is the density of ice, g=9.81 m s^{-2} the acceleration due to gravity, α the surface slope and F is the shape factor (Paterson, 1994):

$$F = \frac{S}{HP} \tag{7}$$

where P is the wetted perimeter of the cross-section.

Eq. (4) is solved on a staggered finite difference grid using a forward explicit time step. The time step is controlled by the stability condition:

$$\Delta t \le \frac{\Delta x^2}{2n\max D} \tag{8}$$

where x is oriented in the direction of flow, Δx is the grid spacing and

$$D = \frac{HU}{\sin\alpha} \tag{9}$$

is the effective diffusion (Hindmarsh, 2001). The time step Δt changes with changing glacier geometry and is typically 0.02 yr.

3.2. Model configuration

The accumulation area of the Franz Josef Glacier is comprised of three major basins, and another three glaciers join or nearly join the glacier further downstream near 1600 m a.s.l. In order to account for complexity such as this, a common scheme is to use more than one flowline (Greuell, 1992; Zuo and Oerlemans, 1997), and this approach is followed here.

For Franz Josef Glacier, six flowlines are defined, as shown in Fig. 1. The boundary condition at the head of each of these flowlines is zero thickness where the flowline starts on steep mountain sides. The flowline for the Geikie Snowfield has a zero flux condition at the head as it begins at an ice divide. Flux in the last segment of each of the subsidiary glacier flowlines is converted to a change in ice thickness and distributed to the adjacent grid points in the main flowline.

3.3. Input data

The input data required for the model described above are the glacier geometry and mass balance. Glacier geometry is defined by the surface elevation h, the bedrock elevation b, the glacier width w_s and w_b, and the valley side slopes θ (Fig. 2).

3.3.1. Surface elevation

Surface elevation data are available in the form of a digital elevation model (DEM) prepared from a contour map which was drawn from aerial photography in 1986 (NZMS 261, G35 and H35). The vertical accuracy of the DEM grid is 25 m, and the horizontal resolution is 20 m. Further surface elevation data are available from GPS profiles near the head and terminus of the glacier recorded during the period 2000–2003. The surface elevation h is calculated from this DEM at each of the flowline grid points (Fig. 1).

3.3.2. Bedrock elevation

Bedrock elevation was measured over the accessible parts of the glacier during January to July 2002, and March 2003, using radio-echo sounding (RES) as described in Anderson (2004). For parts of the glacier where it is not possible to measure bedrock elevation, it was calculated by assuming a constant τ=150 kPa in Eq. (6). Bedrock elevation is calculated for each of the grid points in this way (Fig. 3).

3.3.3. Glacier width and side slopes

Glacier width is calculated using the DEM in conjunction with the glacier outline from the same source (NZMS 261, G35 and H35). In order to preserve the area–elevation relationship of the glacier, the width is calculated as the length of the contour passing through each grid point. Valley side slope θ is calculated from the elevation change over a distance of 100 m outside the glacier outline. The mean slope of the two sides of the glacier is the valley side slope θ.

3.3.4. Mass balance

The mass balance B is required for the model at each grid point and each time step. The mass balance is calculated using a degree–day model which has been constructed for the Franz Josef Glacier using climate data measured on the glacier for three years, and verified using mass balance measurements over three years. The model is explained in detail by Anderson et al. (2006).

3.3.5. Model tuning

In addition to the input data detailed above, the flow parameters f_d and f_s also have to be determined for the model. Present-day (2000–2002) glacier velocity has been measured by Anderson (2004) and flow parameters tuned to fit the velocity distribution. A very good match between modelled and measured ice velocity is found when the flow parameters are set to f_d=1.5×10^{-24} Pa^{-3} s^{-1} and f_s=4.5×10^{-20} Pa^{-3} m^2 m^2 s^{-1}.

3.4. Model characteristics

3.4.1. Length and volume sensitivity

For temperature changes the length sensitivity of the model is 2.7 km °C^{-1} (mean for ±1 °C change) and volume sensitivity 1.6 km^{3}km^3 °C^{-1}. Mean precipitation sensitivity is 1.4 km change in glacier length for ±10%, and 0.4 km^3 change in glacier volume. The precipitation and temperature sensitivity of the model are plotted by Anderson and Mackintosh (2006). Oerlemans (1997a) calculated a length sensitivity for temperature changes of 2.4 km °C^{-1} (mean for ±1 °C change) and a volume sensitivity of 0.9 km^3 and precipitation sensitivity of 1.0 km length change for a ±10% precipitation change. Hence the model appears

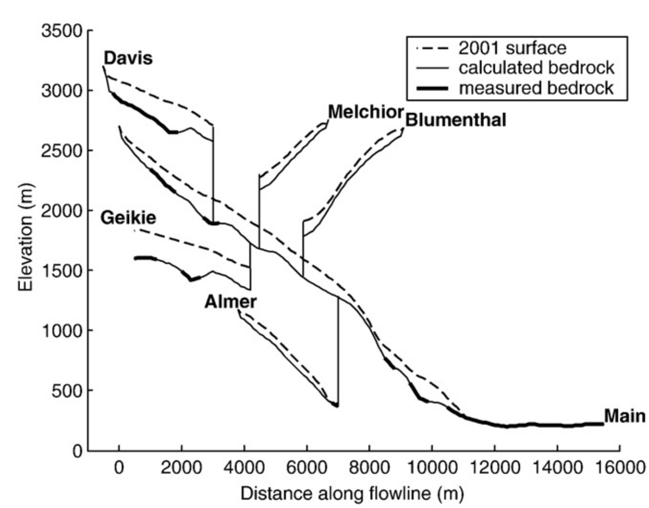

Fig. 3. Surface and bed elevation of the main and tributary flowlines. The plotted elevation of the tributary flowlines has been adjusted for clarity, and the point of connection with the main flowline shown with a vertical line.

to be somewhat more sensitive than the model of Oerlemans (1997a), especially with relation to volume changes.

3.4.2. Response time

The response time of the model, defined here as the e-folding response time (Oerlemans 2001), is calculated by imposing a step change in mass balance and measuring how long the model takes to complete two-thirds of its total response. For a +0.5 °C temperature change, the response time is calculated at 24 yr, and for a −0.5 °C change a response time of 21 yr. These results are similar response to the 20 yr and 27 yr calculated by Oerlemans (1997a) for the same forcing. The volume response times are also very similar.

4. Simulating past glacier response

In order to test the coupled model, a simulation of the terminus position of the glacier is attempted. The flow model is driven with a reconstructed mass balance record (Anderson et al., 2006) and the resulting modelled terminus position compared with the measured record. The terminus position record starts in 1867 (Chinn, 1989; Ruddell, 1995), and is simulated from 1894 to the present. As a first step in this test, a dynamic calibration is performed which ensures that the glacier is in the correct dynamic state in 1894 before the mass balance forcing is applied (Oerlemans, 1997b).

The coupled glacier model run overestimates the glacier length for most of the simulation period (Fig. 4). Despite the dynamic calibration procedure, the model glacier rapidly advances 900 m from its 1894 position, while the measured glacier length stays within 100 m of 13.1 km until 1935. The modelled glacier then retreats almost in parallel with the measured glacier converging slowly until 1985 when the measured glacier advance cuts across the model retreat.

The Hokitika climate record is a composite record made from a number of shorter records around the Hokitika area. The record from 1894 to 2004 has three corrections applied, −1.4 °C from 1894 to 1911, −0.3 °C from 1911 to 1945 and +0.3 °C from 1945 to 1963. These corrections were developed by Salinger (1981) and are outlined in Gellatly and Norton (1984). We find that if the corrections are adjusted in the early part of the record, to −0.9 °C from 1984 to 1911, and the correction from 1911 to 1945 removed, the coupled model simulates glacier length variations much more closely (Fig. 4).

With the corrections adjusted, some features of the coupled model become obvious. The glacier model simulates the general pattern of 20th century advance and retreat very well. However, it is apparent that the modelled glacier length lags the measured glacier length by about 5 yr. The model also does not simulate the peaks of the advance and retreat cycles — the advance in the 1940s is simulated, but the glacier model does not simulate the magnitude of the recession before the advance, nor the full extent of the advance. The advance in the 1960s is not simulated, although the retreat stops. The advance in the 1980s is simulated but, as with the 1940s advance, the magnitude of the recession before the advance is not reached. However, the modelled advance reached at least as far as the measured advance by 2004.

The two model runs, one with the standard corrections to the Hokitika data, and the other with adjusted corrections, are identical from 1994 to 2004 as the effect of the earlier differences in the climate record has been 'forgotten' by the glacier by this time. In both cases the most recent advance came to an end in 2004 in both of the model runs. By 2004 the measured glacier had retreated 500 m in the last five years.

While the simulation is not perfect, the modelled and measured glacier lengths are very close, especially for the last 20 yr of the simulation. This provides some confidence that the mass balance and dynamics of the glacier are being simulated adequately and provide a robust basis for estimating future glacier change.

5. Future prediction

Predicting the future is an uncertain exercise, and dealing with this uncertainty is a central issue in making future predictions of glacier advance and retreat. Ideally, one would construct a probability density

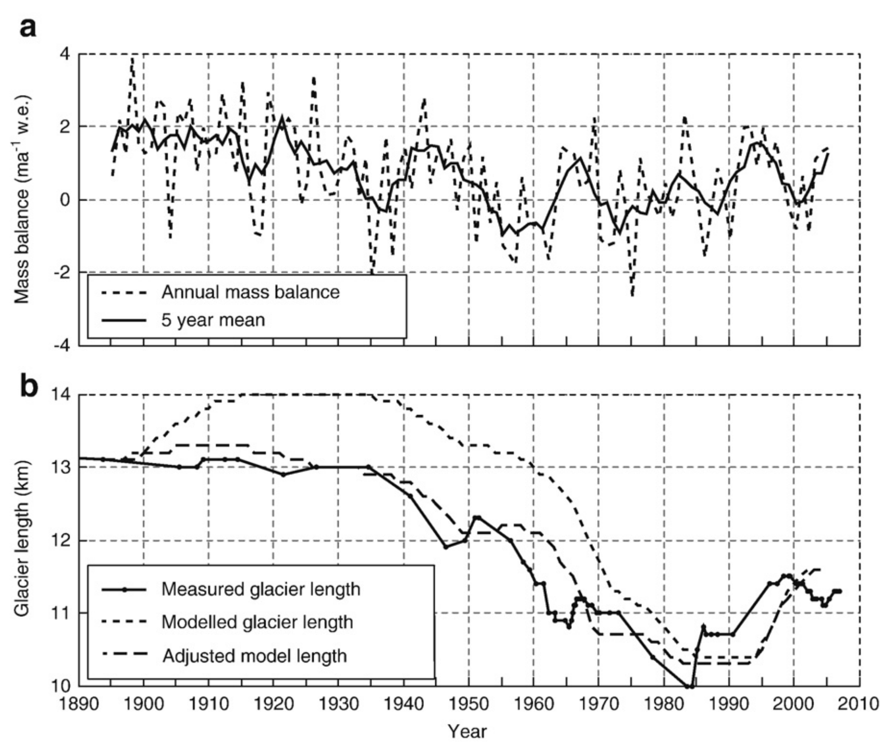

Fig. 4. (a) A mass balance reconstruction, relative to the 1986 glacier geometry, calculated from Hokitika climate data (Anderson et al., 2006) and used to drive the flow model. (b) Glacier length. The glacier length record from 1894 to 1990 used here has been compiled by Ruddell (1995) and since 1996 GPS measurements of the terminus position have been made by the authors.

function which indicates the probability of any particular outcome occurring (van der Veen, 2002).

Here, the future glacier behaviour projections are based on future climate projections. The uncertainty in future climate projections is managed by the use of scenarios, which cover a range of possibilities, but give no indication of the probability of any particular scenario occurring. Because of this limitation in the input data, a probability density function of future glacier behaviour cannot be constructed (van der Veen, 2002), and so a series of glacier scenarios is created. The time frame used for future simulation is that of the IPCC climate change scenarios (Houghton et al., 2001) which provide projections to 2100.

Previous studies of glacier response have used two main approaches to determining future climate characteristics. Initially, global temperature changes associated with different scenarios were applied sometimes with precipitation changes as well (e.g. Oerlemans 1997b; Oerlemans et al., 1998). More recently, temperature and precipitation changes in the local area of interest have been obtained by transient general circulation models (GCMs). In this study, the second approach was adopted following the work of Mullan et al. (2001) who examined output from several transient atmosphere–ocean GCMs, discarded the extremes to give mean, maximum and minimum predictions for temperature, precipitation and pressure changes over the New Zealand region to 2100. They then downscaled these to determine local temperature and precipitation changes using statistical relationships which account for the effect of topography under a changed regional circulation.

Table 1
Temperature and precipitation changes for future climate scenarios used to drive the glacier model

Scenario	2050		2100	
	Temperature (°C)	Precipitation (%)	Temperature (°C)	Precipitation (%)
1. No change	0	0	0	0
2. Minimum change	+0.5	+2.4	+0.9	+4.8
3. Mean change	+0.7	+3.7	+1.4	+8.0
4. Maximum change	+1.1	+5.6	+2.1	+11.1
5. 0.02 K a^{-1}	+1.0	0	+2.0	0

These downscaled scenarios have been used to examine the effect of climate change on the mass balance at Franz Josef Glacier (Anderson et al., 2006). This mass balance variation, calculated for the 1986 geometry and termed the 'reference mass balance', is shown in Fig. 5a for five scenarios:

1. the 'no change' scenario where temperature and rainfall are kept at their means for the 1970–1999 period

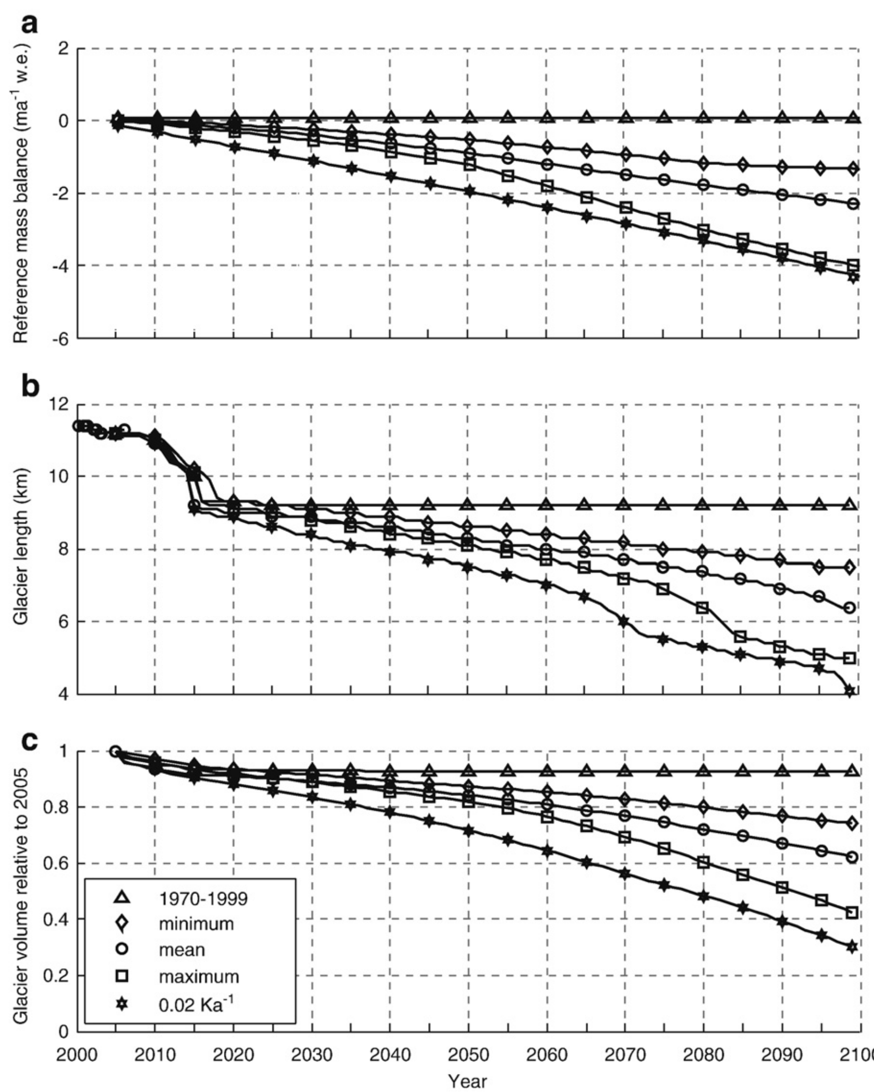

Fig. 5. (a) The mass balance in the upper pane resulting from each of the five climate scenarios is used to drive the glacier flow model; (b) the resulting glacier length; and (c) the resulting glacier volume.

2. the 'mean change' scenario which is the mean global warming scenario, mean monthly temperature change and mean monthly precipitation change
3. the 'minimum change' scenario which is the minimum global warming scenario, mean monthly temperature change and mean monthly precipitation change
4. the 'maximum change' scenario which is the maximum global warming scenario, mean monthly temperature change and mean monthly precipitation change
5. the '0.02 K a^{-1} warming' scenario in which the temperature is increased by 0.02 K per annum. This scenario is included to allow comparison with earlier work on Franz Josef and other glaciers (Oerlemans et al., 1998).

The glacier flow model is run with these future climate scenarios, with the particular values of temperature and precipitation change as shown in Table 1. A dynamic calibration from 1980 to 2004 was used to ensure that the glacier was in the correct dynamic state at the start of the model runs, rather than the state at the end of the coupled run where the terminus position was not simulated precisely. The resulting glacier length and volume changes are shown in Fig. 5b and c with changes to 2100 summarised in Table 1.

For all scenarios the glacier retreats for a few years after 2004. This retreat is caused by the dynamic state of the glacier at the start of the future simulation. The retreat continues over two bedrock steps (Fig. 3) to a stable length of approximately 9 km by 2015. After this time the different scenarios diverge rapidly as the long term mass balance signal from the climate change scenarios become important.

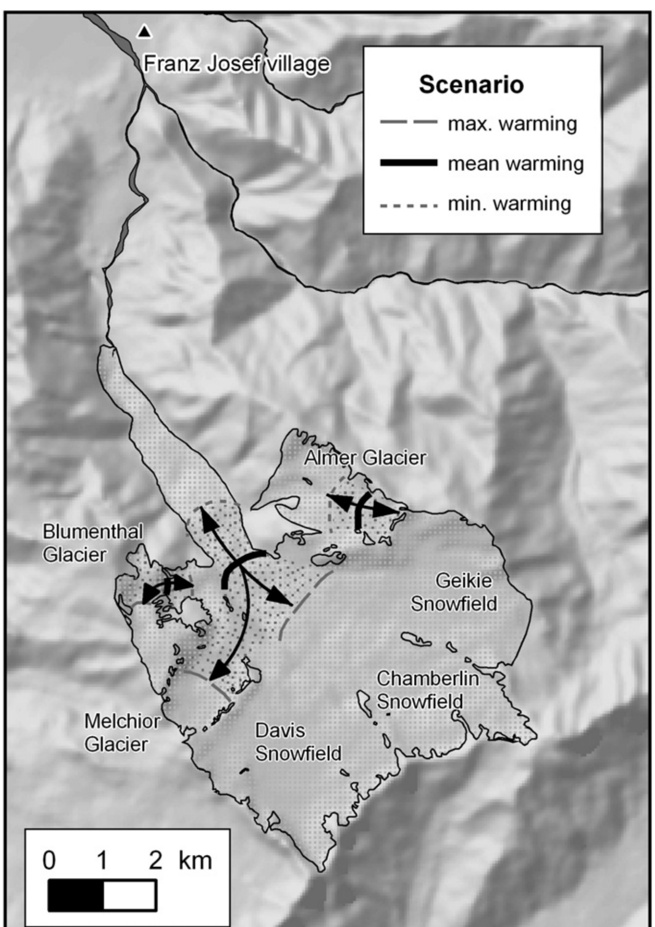

Fig. 6. The approximate terminus position of the glacier in 2100 under different scenarios. Arrows indicate the range of terminus position resulting from the minimum, mean and maximum scenarios.

The changes in the length of the main flowline and of the tributary glaciers which disconnect from the main flowline by 2100 are shown for three scenarios in Fig. 6. The 'mean change' scenario forces a continued slow recession to a length of 6.4 km by 2100, and a reduction in volume to 62% of the 2004 volume. The Almer Glacier retreats to become a small cirque glacier, and the Blumenthal Glacier disconnects from the main flowline in 2072. For the 'maximum change' scenario there is a retreat to 5.0 km in glacier length, and 42% of the 2004 volume. The Blumenthal Glacier disconnects from the main flowline in 2061, and the Melchior Glacier in 2082. The Almer Glacier disappears by the end of the simulation. The 'minimum change' scenario results in a length of 7.5 km, and a reduction in volume to 74% of the 2004 volume. The Blumenthal Glacier disconnects from the main flowline in 2091 and the Almer Glacier reduces to 40% of its present-day length.

The 'no change' scenario shows a stable length of 9.2 km, 2 km shorter than present, close to the 1986 length of 10 km. The 0.02 K a^{-1} scenario results in a reduction in length to 4 km by 2100, and a reduction in volume to 31% of the 2004 volume. This corresponds to a 3.6 km K^{-1} length sensitivity to temperature increase.

6. Discussion

6.1. Past response

Attempts to use mass balance reconstructions to drive glacier flow models have been made by a number of authors, with varying success. Studies which use a long term climate reconstruction to calculate mass balance have generally shown a poor correspondence between measured and modelled glacier terminus changes (Oerlemans, 1986; Stroeven et al., 1989; Huybrechts et al., 1989). Simulations over shorter periods, using mass balance calculated from instrumental climate data have shown a better correspondence (Schlosser, 1997). Schmeits and Oerlemans (1997) succeeded in using a partially reconstructed climate record to drive a glacier flow model which simulated the historic variations of Unterer Gridelwaldgletchser.

In this regard, this study of the Franz Josef Glacier is somewhat unique insofar as it has attempted to simulate response during the period covered by the instrumental climate record which includes four cycles of retreat and advance. While the coupled model used here did not reproduce exact glacier length throughout the simulation, the pattern of advance and retreat is clear. It is apparent that there are problems in both the climate record/mass balance model, which together overestimate mass balance early in the simulation period, and in the glacier flow model, in the timing of advance and retreat is delayed compared with the observed terminus position.

The failure of the standard climate record to simulate the early part of the period maybe due to poor quality climate data in the early part of the record, or some shift in the relationship between Hokitika climate data and Franz Josef Glacier mass balance. The mass balance model was constructed and tested using mass balance and climate data from the period 2000–2005 (Anderson et al., 2006) and it is not possible to distinguish between these two possibilities.

The slow response of the modelled glacier to periods of high or low mass balance compared to observations of glacier length is also apparent in the model output. The model results presented here do not incorporate the influence of longitudinal stress gradients on glacier dynamics. Some studies have found that incorporating longitudinal stress gradients is important in modelling glacier dynamics (Hubbard, 2000), and in the dynamic response to climatic changes (Greuell, 1992). For steady-state simulations the uncertainties in mass balance are more important than uncertainties in ice flow (Greuell, 1992), and Leysinger Vieli and Gudmundsson (2004) emphasise that uncertainty in mass balance is more important that flow model assumptions in assessing dynamic response. However in this case is apparent that the glacier dynamics have not been adequately accounted for. Whether this is a result of using a simple model which neglects longitudinal stress gradients, or other simplifications such as a sliding law which does not include subglacial hydrology is a question that requires further analysis.

6.2. Future response

Long term (100 yr) glacier length and volume predictions must be approached with some caution. Uncertainties in future climate projections are compounded by uncertainties in the mass balance model and uncertainties in the flow model before we arrive at projections of glacier length and volume. However we can provide a range of possibilities for future glacier behaviour.

The range of possibilities of glacier length change all cause the recession of the long, steep tongue of the glacier which reaches to ~300 m a.s.l. and is one of the defining features of the glacier. However, while the length response of the glacier is large, a reduction of 44%, the volume response is milder at 38%. The long, steep tongue acts as a buffer — the glacier can easily retreat to higher ground while losing a significant, but not catastrophic, proportion of its volume.

Compared to other glaciers in New Zealand, the glacier is rather well placed to withstand a warming climate. Due to New Zealand's maritime climate, the mass balance sensitivities of Franz Josef Glacier (Anderson et al., 2006) and the Brewster Glacier (Anderson et al., submitted for publication), 90 km to the south, are both very high (≥1.9 m a^{-1} w.e. K^{-1}). The mean change scenario produces an equilibrium line altitude on Franz Josef Glacier of 2040 m a.s.l. which, based on the glacier inventory of Chinn (1989), is above the maximum elevation of 49% of New Zealand's 3150 glaciers. Consequently, this change is likely to cause the disappearance of many of the glaciers in the Southern Alps.

Comparison of the results of this study with other simulations of future glacier change on a global scale is compromised by the use of different scenarios of climate change and by slightly different time periods. However, some general conclusions can be made from a summary of these studies in Table 2. Oerlemans et al. (1998) showed that smaller glaciers generally can be expected to lose a large proportion of their volume. That study also indicated that Franz Josef Glacier was at the lower end of the range of relative volume losses, at 33% (using a 0.02 Ka^{-1} warming). However, the result from this study for the 0.02 K a^{-1} scenario produces a 69% volume loss, more in line with other mountain/valley glaciers (<100 km^2) which all lose >50% volume under this scenario. The difference between estimates for volume loss at Franz Josef Glacier for the same forcing is partially due to the differing sensitivities of the models. Oerlemans et al. (1998) use an energy balance model. Degree–day models tend to over-state the mass balance sensitivity to temperature changes because they rely on an implicit correlation with solar radiation (Oerlemans, 2001), although Braithwaite et al. (2002) find good agreement between sensitivities calculated from energy balance and degree–day models. However, there is also a difference in the initial state of the glacier used by Oerlemans et al. (1998) who started with a glacier model which grew by 2 km from its 1990 state with no forcing applied (using 1961–1990 climate means). Hence any warming imposed on that model would be offset by the glacier's tendency to grow, and Oerlemans et al. (1998) probably underestimate volume change.

The high sensitivity of the Franz Josef Glacier is offset by its large elevation range so the same warming does not lead to the drastic volume loss of Nigardsbreen. The length change of the Franz Josef glacier is at the upper end of the range for similar-sized mountain glaciers because of the glacier geometry with a long steep tongue.

7. Conclusion

A coupled mass balance and glacier flow model has simulated the general retreat of the Franz Josef Glacier in the 20th Century. While the simulation is not perfect, it indicates that the model system is

Table 2
Year 2100 model projections of final area, volume and length change for 18 glaciers around the world, based on global change rates (unless otherwise noted under comments)

Glacier	Area (km^2)	Local change		Glacier change		Comments	Source
		Temperature (K a^{-1})	Precipitation (%K^{-1})	Volume (%)	Length (km)		
Nigardsbreen	48	0.02		−90	−8		Oerlemans (1997b)
		0.02	10	−58			
Pasterze	19.8	0.028		−71	−5	Downscaled to IPCC IS92a	Zuo and Oerlemans (1997)
		0.028	10	−40			
Rhonegletscher	17.1	0.02		−66	−4		Wallinga and van den Wal (1998)
		0.02	10	−40			
Blöndujökull/Kvislajökull	226	0.03	5	−37	−2		Jóhannesson (1997)
Illviðrajökull	116	0.03	5	−34	−1.5		Jóhannesson (1997)
Unterer Grindelwaldgletscher	21.7	0.02		−66	−2.7		Schmeits and Oerlemans (1997)
Haut Glac. D'Arolla	6.3	0.02		−98		Volume changes estimated from Fig. 3	Oerlemans et al. (1998)
Nigardsbreen	48	0.02		−91			
Hintereisferner	7.4	0.02		−81			
Storglaciären	3.1	0.02		−77			
Rhonegletscher	17.7	0.02		−65			
Pasterze	19.8	0.02		−55			
Unt. Grindelwaldgl.	21.7	0.02		−53			
Gl. D'Argentière	15.6	0.02		−53			
Blöndujökull	226	0.02		−37			
Illviðrajökull	116	0.02		−37			
King George Island	1402	0.02		−34			
Franz Josef	35	0.02		−33			
Average of glaciers in Oerlemans and others(1998)		0.02		−60		Averaged by glacier	Oerlemans (2001)
		0.02	10	−43			
		0.02		−37		Averaged by volume	
		0.02	10	−25			
Franz Josef	35	0.014	5.3	−38	−5.0	Mean downscaled	This study (2005 – 2100)
		0.021	5.3	−58	−6.4	Max downscaled	
		0.009	5.3	−26	−3.9	Min downscaled	
		0	0	−8	−2.2	No change	
		0.02	0	−69	−7.4	0.02 K a^{-1}	

capable of simulating the general patterns of advance and retreat of the glacier, and that the climate signal responsible for these patterns is contained within the Hokitika climate record. Combined with the findings of Anderson et al. (2006), who demonstrate that the general negative trend in mass for this period is caused by increasing temperatures, we can conclude that the general retreat of the glacier is caused by a regional warming.

If the climate changes as expected by the IPCC (Houghton et al., 2001) and downscaled to New Zealand (Mullan et al., 2001), the Franz Josef Glacier will retreat approximately 5 km, removing all of the steep glacier tongue. The volume change will be towards the lower end of the range of predictions for other glaciers, with a 38% reduction. However, the 5 km retreat will result in a dramatic change to the nature of the glacier. While the consequences for tourism at this glacier may be serious, the large elevation range of Franz Josef Glacier means that it will withstand the projected warming better than most New Zealand glaciers.

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