Cloud effects on the surface energy and mass balance of Brewster Glacier, New Zealand

J. P. Conway¹,² and N. J. Cullen¹

¹Department of Geography, University of Otago, Dunedin, New Zealand
²Centre for Hydrology, University of Saskatchewan, Saskatoon, Canada

Received: 19 January 2015 – Accepted: 31 January 2015 – Published: 18 February 2015

Correspondence to: J. P. Conway (jonathan.conway@usask.ca)

Published by Copernicus Publications on behalf of the European Geosciences Union.
Abstract

A thorough understanding of the influence of clouds on glacier surface energy balance (SEB) and surface mass balance (SMB) is critical for forward and backward modelling of glacier–climate interactions. A validated 22 month time series of SEB/SMB was constructed for the ablation zone of the Brewster Glacier, using high quality radiation data to carefully evaluate SEB terms and define clear-sky and overcast conditions. A fundamental change in glacier SEB in cloudy conditions was driven by increased effective sky emissivity and surface vapour pressure, rather than the minimal change in air temperature and wind speed. During overcast conditions, positive net longwave radiation and latent heat fluxes allowed melt to be maintained through a much greater length of time compared to clear-sky conditions, and led to similar melt in each sky condition. The sensitivity of SMB to changes in air temperature was greatly enhanced in overcast compared to clear-sky conditions due to more frequent melt and the occurrence of precipitation, which enabled a strong accumulation–albedo feedback. During the spring and autumn seasons, the sensitivity during overcast conditions was strongest. There is a need to include the effects of atmospheric moisture (vapour, cloud and precipitation) on melt processes when modelling glacier–climate interactions.

1 Introduction

The response of glaciers to atmospheric forcing is of interest as glaciers are seen as useful scalable proxy records of past climate (e.g. Mölg et al., 2009a) and because the rapid changes occurring in many glaciated regions have implications for both global sea level rise (Kaser et al., 2006) and water resources (e.g. Jost et al., 2012). Reliable attribution of past glacier states and prediction of future ones is dependent on a thorough understanding of the physical processes operating at the glacier surface that link glacier change with climate, that is, the surface mass balance (SMB) and surface energy balance (SEB). For debris free, mid-latitude glaciers, the SMB is primarily a prod-
uct of the relative magnitudes of accumulated solid precipitation and melt. While, in
general, incoming shortwave radiation (SW↓) is the major source of energy for glacier
melt, variations in SMB are considered to be forced by changes in air temperature
and precipitation (Oerlemans, 2005) through both accumulation and melt processes.
A strong positive feedback between accumulation and surface albedo accounts for
much of the sensitivity to both air temperature and precipitation, along with the often
efficient relationship between air temperature and melt. The mechanisms responsible
for the temperature dependence of melt vary widely (Sicart et al., 2008), and include
the variability of turbulent sensible (QS) and latent (QL) heat fluxes, incoming longwave
radiation (LW↓), and a (somewhat spurious) covariance between air temperature and
SW↓ in many continental areas. The primary influence of air temperature on melt rate
is also nuanced by other influences on the SEB such as surface albedo (Oerlemans
et al., 2009), humidity (Gillett and Cullen, 2011), and cloud transmission (Pellicciotti
et al., 2005).

The influence of clouds on the SEB is, in fact, far more pervasive. Recent advances in
AWS deployment on glacier surfaces (Mölg et al., 2009b), the availability of high-quality
radiation measurements (van den Broeke et al., 2004), and development of methods
to extract information about cloud cover in data sparse areas (Kuipers Munneke et al.,
2011), have allowed the variation of SEB and SMB with cloud cover to be characterised
in many areas. Sicart et al. (2010) show clouds dominate day to day variations in LW↓
in mountainous areas while numerous studies detail the fundamental changes in SEB
with cloudiness that are often co-incident with changes in glacier surface boundary
layer (SBL) properties (van den Broeke et al., 2006; Giesen et al., 2008; Gillett and
Cullen, 2011). Given their strong control on the SEB, and coincidence with changes
in SBL properties it is vital that the role of clouds in altering the sensitivity of SMB to
changes in atmospheric state variables (especially air temperature) be assessed.

The glaciers of the Southern Alps of New Zealand occupy a unique position in the
westerly wind belt of the Southern Ocean, a region dominated by mid-latitude atmo-
spheric circulation (Tait and Fitzharris, 1998; Ummenhofer and England, 2007). The
large barrier the Southern Alps poses to the prevailing winds creates a high precipita-
tion environment, which, coupled to the relatively low elevation of glacier termini
(Hoelzle et al., 2007), creates high mass turnover glaciers that have shown high sen-
sitivity to climatic variations in temperature-index glacier modelling studies (Anderson
et al., 2006; Oerlemans, 2010). For these reasons the glaciers of the Southern Alps
are seen as useful indicators of regional atmospheric circulation in the southwest Pa-
cific and form a vital component of paleoclimate work (e.g. Lorrey et al., 2007). While
accumulation–albedo feedbacks have been shown to be important to the sensitivity
of SMB to air temperature in New Zealand as in other glaciated regions (Oerlemans,
1997; Anderson et al., 2006), there is a suggestion that increased turbulent (mainly
sensible) heat fluxes dominate variations in melt (Anderson et al., 2010). This has
led some authors to interpret past glacier fluctuations as a linear and direct proxy for
regional air temperature (e.g. Putnam et al., 2012), at the exclusion of most other ele-
ments of the glacier–climate system.

It has been well established that synoptic scale processes exert a strong control on
the SMB in the Southern Alps, with periods of 20th Century glacier advance and re-
treat associated with anomalies in the regional climate system (Fitzharris et al., 2007).
Given that this synoptic variability is closely linked to inferred changes in cloudiness as
well as airmass properties (Hay and Fitzharris, 1988), and that these synoptic controls
are thought to have varied over paleo-climatic timescales (Drost et al., 2007; Ackerley
et al., 2011), it is vital that the influence of clouds on SMB is separated out from the
influence of airmass properties (in particular air temperature). Recent field studies on
the Brewster Glacier, Southern Alps, have shown the high frequency of cloudy condi-
tions during all seasons (> 50 % overcast conditions) as well as the significant and
variable effect of clouds on SW↓, LW↓ and net radiation (Rnet) (Conway et al., 2014).
In this context it is timely to examine in detail the influence of clouds on glacier surface
climate, SEB and melt, as well as the manner in which clouds alter the sensitivity of
SMB to air temperature in the Southern Alps.
This paper addresses these issues by resolving the SEB and SMB at a site in the ablation zone of the Brewster Glacier over a 22 month period. High quality surface climate data presented in Cullen and Conway (2015) are used to force a SMB model (Mölg et al., 2008) to estimate both SEB and SMB terms over this period (measurement period). The cloud metrics presented in Conway et al. (2014) are used to identify clear-sky and overcast conditions and thus characterise surface climate, SEB and melt energy during each condition. To test the sensitivity of SMB to changes in surface climate and radiative components, a more heavily parameterised version of the model is run over a hybrid two-year dataset (sensitivity period), allowing the effect of changes in surface climate and radiative properties to be assessed independently and the influence of clouds on this sensitivity to be assessed. The following section provides a brief description of the site, datasets and modelling methods before the results and discussion are presented in subsequent sections.

2 Methods

2.1 Site description and instrumentation

The Brewster Glacier is a small mountain glacier situated in the Southern Alps immediately west of the main divide (Fig. 1). It experiences a temperate maritime high precipitation environment, with annual precipitation around 6000 mm water equivalent (w.e.) and an annual temperature of 1.2°C over the glacier surface at 1760 m a.s.l. (Cullen and Conway, 2015). In comparison to other glaciers in the Southern Alps, it has a somewhat lower average slope (16°) but similar mean elevation and elevation of the glacier snout (Hoelzle et al., 2007). As it is located on the main divide with relatively high exposure to synoptic weather systems, at the midpoint of the north–south distribution of glaciers in the Southern Alps (Chinn et al., 2012), it is likely to experience the atmospheric controls on SMB that affect the Southern Alps in general.
Data from an automatic weather station (AWS) situated in the ablation area of Brewster Glacier (AWS\textsubscript{glacier}) were used in this study (Fig. 1). Table 1 gives details of instrumentation and annual average surface climate variables at AWS\textsubscript{glacier}, while further details of the locality and AWS instrumentation can be found in Cullen and Conway (2015). Measurements at AWS\textsubscript{glacier} ran for 22 months from 25 October 2010 to 1 September 2012 (inclusive). Air temperature ($T_a$) shows a moderate seasonal cycle (8°C), and airmass changes appear to override the subdued diurnal range in $T_a$. Wind speed ($U$) is moderate with a persistent down-glacier flow despite the small fetch and exposed location (Conway, 2013). Humidity is high with average vapour pressure exceeding that of a melting surface through 4 months during summer. Cloud cover is frequent and associated with on-glacier wind direction (Conway et al., 2014). Annual mass balance in the vicinity of AWS\textsubscript{glacier} is generally negative, despite the large accumulation (> 3 m.w.e.) of winter snowfall during May through September. The significant annual ablation (> 4 m.w.e.) generally starts during October, exposing an ice surface in early January and continuing till April or later.

2.2 Data treatment and cloud metrics

Cullen and Conway (2015) describe the treatment of the AWS data in detail but a summary of the main steps is given here. $T_a$ data were corrected for the (large) influence of solar radiation on the unaspirated shields, resulting in a 0.7°C decrease in mean annual $T_a$. To facilitate SMB modelling, a continuous precipitation dataset ($P_{scaled}$) was constructed by comparing summer rain gauge observations from a second AWS situated in the pro-glacial area (AWS\textsubscript{lake}) to a nearby lowland rain gauge ($R^2 = 0.9$ at daily level).

To construct a high temporal resolution record of observed SMB, surface height observed using a sonic ranger (Cullen and Conway, 2015) was combined with periodic snow density measurements. Snow pits near the start of snowmelt indicated a consistent density approaching 500 kg m$^{-3}$ during late October (443 kg m$^{-3}$ on 23 October 2010; 483 kg m$^{-3}$ on 27 October 2011), while density during mid-winter was more
moderate (320 kg m\(^{-3}\) on 18 July 2011). Thus, while the density of melting snow during spring is relatively well constrained, the increasing density due to subsurface processes (e.g. viscous compaction and melt–refreezing) during the winter months produces some uncertainty in the relationship between surface height and SMB. Beyond the snow-ice transition in early January, a standard ice density of 900 kg m\(^{-3}\) was assumed, while short periods of new snowfall were assigned a fresh snow density of 300 kg m\(^{-3}\) (Gillett and Cullen, 2011).

A record of cloudiness was constructed using measurements of LW\(_\downarrow\) and a clear-sky emissivity model (Conway et al., 2014). The longwave equivalent cloudiness (\(N_\varepsilon\)) scales the effective sky emissivity (from observed LW\(_\downarrow\)) between the modelled clear-sky emissivity and an overcast emissivity of 1. Though not directly comparable to traditional cloud fraction metrics based on manual or sky camera observations, \(N_\varepsilon\) very effectively characterises the effects of clouds on surface radiation fluxes. It also has the advantage over metrics based on SW\(_\downarrow\), in that it provides 24 h coverage and is not affected by solar zenith angle or multiple reflections between the surface and atmosphere.

### 2.3 Model description

A SMB model (Mögl et al., 2008) was used to resolve surface energy and mass fluxes at AWS\(_{\text{glacier}}\) for the full 22 month study period. A full description of the model is given in Mögl et al. (2008, 2009a), but a short description of the parameterisation of each term is given here. The model computes SMB as the sum of snow accumulation, melt, refreezing of liquid water in the snowpack and mass fluxes of water vapour (deposition and sublimation) while surface temperature (\(T_s\)) is less than 0°C. Fluxes of vapour while the surface temperature is melting are not included directly in the SMB, as condensation and evaporation are assumed to add and remove mass from the liquid melt water at the surface, respectively. The model uses \(T_s\) as a free variable to close the SEB (Eq. 1) at each 30 min timestep:

\[
QM = SW\downarrow (1 - \alpha) + LW\downarrow - \sigma \varepsilon T_s^4 + QS + QL + QR + QC
\]

\[\text{(1)}\]
where \( Q_M \) is the energy for surface melt while \( T_s = 0 \)°C, \( SW_\downarrow \) is the incoming solar radiation, \( \alpha \) is the albedo, \( LW_\downarrow \) is the incoming longwave radiation, \( \sigma \) is the Stefan–Boltzmann constant (\( 5.67 \times 10^{-8} \) W m\(^{-2}\)), \( \varepsilon \) is the emissivity of snow/ice (equal to unity), \( T_s \) is the surface temperature (K), \( QS \) and \( QL \) are the turbulent sensible and latent heat fluxes, respectively, \( QR \) is the rain heat flux and \( QC \) is the conductive heat flux through the glacier subsurface. The convention used is that energy fluxes directed towards the surface are positive.

Two different configurations of the model are presented in this paper, distinguished only by their treatment of surface radiation fluxes. For the first, SEBmr, we used measured values of \( SW_\downarrow \), \( LW_\downarrow \) and \( \alpha \) from AWS\(_{\text{glacier}}\) (Table 2) to provide best estimates of SEB and SMB terms for analysis. For the second, SEBpr, we used parameterised radiation fluxes (Table 2) to assess the sensitivity of the SMB to changes in surface climate (detailed further in Sect. 2.5). All other energy fluxes are calculated consistently between configurations. \( QR \) is calculated using \( P_{\text{scaled}} \) assuming rain temperature is equal to \( T_a \). New snow was calculated from \( P_{\text{scaled}} \) using a rain/snow threshold (\( T_{r/s} \)) of 1 °C and a fixed density of 300 kg m\(^{-3}\). The iterative SEB closure scheme of Mölg et al. (2008) was used to calculate \( T_s \), with QC being calculated as the flux between the surface and the top layer of the twelve layer subsurface module (subsurface levels: 0.1, 0.2, 0.3, 0.4, 0.5, 0.8, 1.4, 2, 3, 5, and 7 m). Penetrating shortwave radiation was not included in the model, as the subsurface temperature profile was not measured throughout the study period; hence the optimisation of a penetrating shortwave radiation scheme would be subject to large uncertainty. The depth, density and temperature (iso-thermal at 0 °C) of the snowpack was prescribed at the start of the measurement period from snow-pit measurements (see Sect. 2.2).

The turbulent heat fluxes, \( QS \) and \( QL \), were calculated using a bulk-aerodynamic approach using the \( C_{\text{log}} \) parameterisation as described by Conway and Cullen (2013). The roughness lengths for momentum (\( z_{0v} \)), temperature (\( z_{0t} \)) and humidity (\( z_{0q} \)) over an ice surface at AWS\(_{\text{glacier}}\) are well constrained by in-situ measurements (\( z_{0v} = 3.6 \times 10^{-3} \) m, \( z_{0t} = z_{0q} = 5.5 \times 10^{-5} \) m; Conway and Cullen, 2013), though spatial and
temporal variability is still probable. A further period of eddy covariance measurements over a spring snow surface (27 October to 3 November 2011) showed a log-mean value for $z_{0v}$ of $1.8 \times 10^{-3}$ m ($\sigma = 1.3 \times 10^{-2}$ m, $n = 31$), using the same filtering criterion as Conway and Cullen (2013). No reliable estimates of $z_{0t}$ or $z_{0q}$ were possible because of the large uncertainties involved with the small temperature and vapour pressure gradients experienced during this period. Given the similar, but more uncertain, $z_{0v}$ over snow and the large effect of $z_{0t}$ on the effective roughness length which tends to counter a change in $z_{0v}$ (Conway and Cullen, 2013), roughness lengths derived over ice were adopted for the entire period.

2.4 Estimation of uncertainty using a Monte Carlo approach

To estimate uncertainty in modelled SMB, a series of Monte Carlo simulations were made covering the range of input data and parameter uncertainty expected for each configuration of the model (SEBmr and SEBpr). Table 3 shows the parameter uncertainty introduced for each configuration, while input data uncertainty was kept consistent with that used in Conway and Cullen (2013) and is given in Table 1. For both configuration, 5000 runs of the measurement period were made, with systematic and random errors being assigned to each input variable before each simulation and time step, respectively. Errors were calculated by multiplying the uncertainties associated with each input variable (Tables 1 and 3) by normally distributed random numbers ($\mu = 0; \sigma = 1$), with the exception of $z_{0v}$ which was logarithmically transformed before the uncertainty was applied. The 5000 SMB time series computed for each configuration were subjected to a first order check, using measured $T_s$ as a proxy for a realistic evaluation of the SEB. Runs were removed when 30 min modelled $T_s$ had root mean squared differences (RMSD) $> 1.5$ K or $R^2 < 0.9$, which removed $\sim 10\%$ of runs from each ensemble. The remaining runs were then used to compute an ensemble mean and SD for the SMB accumulated over one-day and 10 day periods in addition to the full measurement period. Runs that did not correctly predict the accumulated SMB at the end of the measurement period were not removed, as it was unknown if any systematic
errors would remain constant over the study period. Thus, the model uncertainty over a shorter time period (e.g. one or 10 days) was kept independent of the final “correct” accumulated SMB.

2.5 Mass balance sensitivity configuration

To assess the mass balance sensitivity ($\Delta$SMB) at AWS$_{\text{glacier}}$ further runs were made with the SEBpr configuration using a hybrid 2 year dataset (sensitivity period). The goal was not only to show the extent to which elements of the climate system could force SMB changes but also to understand how uncertainty in model input data or parameterisation impacted estimates of SMB. Because the measurement period started in spring, the initial depth and density of the snowpack was prescribed in these runs. However, a realistic evolution of snowdepth with perturbations in surface climate (especially $T_a$) is required to assess $\Delta$SMB, i.e. $\Delta$SMB is assessed with accumulation seasons preceding ablation seasons. To this end, a hybrid two-year dataset was constructed using data from AWS$_{\text{glacier}}$ by rearranging the measurement period timeseries. The particular periods used were (in order): 1 May to 1 September 2012, 2 September to 24 October 2011 and 25 October 2010 to 30 April 2012. This gave two full SMB seasons (1 May–30 April) in sensitivity runs and retained variability in the input data without relying on data from off-glacier sources. Fortunately, the snowdepth predicted by SEBpr at the end of the first hybrid accumulation season matches that at the start of the measurement period (25 October 2010) so the evolution of snowdepth (and $\alpha$) during the remainder of the sensitivity run is comparable with that in the measurement period.

To enable snowdepth–$\alpha$ feedbacks within SEBpr, $\alpha$ was simulated using the parameterisation of Oerlemans and Knap (1998). This scheme computes albedo from three values representative of fresh snow, firn and ice, accounting for the evolution of fresh snow to firn through an e-folding constant ($t^*$) which describes the characteristic albedo timescale. Two modifications were made to the scheme (Mölgl et al., 2012). Firstly, when new snowfall is removed by melt, the albedo reverts back to the albedo
of the underlying surface. Secondly, a daily total snowfall in excess of 5 cm (depth) was introduced as a threshold above which the new snowfall impacts albedo, as small snowfalls are likely to be redistributed into crevasses and hollows on the glacier surface and have minimal impact on the albedo. These two modifications removed large positive feedbacks in the scheme that often caused small snowfalls to increase albedo long after the model had removed the new snow by melting.

An analysis of measured albedo $\alpha_{\text{acc}}$ at AWS$_{\text{glacier}}$ allowed local values of $\alpha_{\text{firsnow}}$ (0.95), $\alpha_{\text{firn}}$ (0.65) and $\alpha_{\text{ice}}$ (0.42) to be defined (Fig. 2a). The higher local values are likely indicative of lower levels of contaminants that are responsible for decreased albedo (Oerlemans et al., 2009) and a lack of debris surrounding the Brewster Glacier. A better fit to the evolution of measured albedo was found by decreasing $t^*$ to 10 days, which seems reasonable given the higher rate of melt (and therefore snow metamorphism) in this maritime environment. Figure 2a also shows a marked difference between the ice surface albedo between the two seasons. It is unclear if this difference reflects changes over a large spatial scale or if a localised increase in sediment observed in the vicinity of AWS$_{\text{glacier}}$ during the summer of 2012 contributed to the decrease in albedo during the second season. Without a clear basis for this variation, a mean value of $\alpha_{\text{ice}} = 0.42$ is adopted for both seasons.

$\Delta$SMB was computed by conducting runs with SEBpr over the sensitivity period, introducing a range of systematic perturbations to input data and parameters (introduced with the results in Table 6) and comparing SMB between each run. To calculate variations in $\Delta$SMB with cloudiness, $\Delta$SMB was computed at each model timestep (i.e. mmw.e. 30 min$^{-1}$) for each perturbation run. Model timesteps were then selected based on cloudiness ($N_c$) and a monthly average produced for clear-sky and overcast conditions. For ease of interpretation, $\Delta$SMB was converted to a daily rate (mmw.e.day$^{-1}$) by multiplying by the number of model timesteps within a day (48). By definition, the sum of $\Delta$SMB for each timestep within a year is equal to the accumulated $\Delta$SMB of the entire year, which is the more commonly reported value (e.g. 1.5 mw.e.a$^{-1}$).
3 Results

3.1 Model evaluation

Both configurations of the SMB model (SEBmr and SEBpr) were first validated against observed $T_s$ and SMB during the measurement period. Modelled $T_s$ from reference runs of both configurations agreed well with $T_s$ calculated from measurements of outgoing longwave radiation (Fig. 3). Both mean bias error (MBE) and RMSD at 30 min timesteps were comparable to similar studies (van den Broeke et al., 2011), and monthly averages indicated no seasonally dependant errors in the SEB. Figure 2b shows the large accumulation and ablation experienced at AWS$_{\text{glacier}}$ during the measurement period. Winter accumulations were fairly consistent between 1.5 and 1.8 m.w.e., while summer ablation was more variable. In general, both configurations of the model gave fairly close agreement to the observed accumulated SMB over the measurement period. Both gave SMB during the first accumulation season within $\pm 10\%$ of that observed (Table 4), which was encouraging given the uncertainties in the scaled precipitation dataset and rain/snow threshold. SEBmr showed small discrepancies in modelled ablation (around 10 %) for the ice surface in the first season and the snow surface in the second season (Table 4). SEBpr showed similar performance, with an underestimate of ablation for ice surface in the second season likely related to the lower measured $\alpha$ in this season (Fig. 2a). Despite these small deviations, both configurations produced SMB over the two seasons that was well within the accumulated uncertainty due to measurement and parameter errors (grey shading in Fig. 2b). The small discrepancies between modelled and observed ablation could have been removed, perhaps through specifying different $z_{ov}$ for snow and ice surfaces. However, given that the deviations were not consistent between each season and model, that both models exhibited large accumulated uncertainty, and that our interest was primarily at shorter timescales, we found no strong reasoning for tuning model parameters to fit model values precisely.

To ensure we could correctly simulate the large temporal variability in accumulation and ablation with each configuration of the model we also compared SMB over one-day
and 10 day periods (Fig. 4). SEBmr effectively captured the large variability in SMB during both accumulation and ablation seasons with maximum 10 day ablation and accumulation rates on the order of 50 mm w.e. day\(^{-1}\) (Fig. 4b). A consistent bias in ablation was not observed, confirming our decision not to tune modelled melt exactly over the season. The significant number of large daily ablation events (\(>\) 50 mm w.e. day\(^{-1}\)) observed in the ablation record were, in general, captured by SEBmr (Fig. 4a). If anything, a bias toward under-prediction of these events was seen. This bias is likely related to an under-prediction of QR, as the time-averaging \(P_{\text{scaled}}\) underestimated the very intense rainfall rates (\(>\) 100 mm\(\text{day}^{-1}\)) associated with the largest ablation events (Gillett and Cullen, 2011). 10 day accumulation rates were captured well while daily totals exhibited much larger scatter, reflecting the difficulty of determining observed winter SMB from surface height records as well as the large combined uncertainty due to \(P_{\text{scaled}}\), \(T_a\) and \(T_{r/s}\). The good agreement of modelled and observed SMB at these short temporal resolutions suggests SEBmr is able to capture the variations in melt and accumulation forced by the key synoptic atmospheric controls.

SEBpr showed similar agreement to observed SMB at both daily and 10 day level (Fig. 4c and d). The larger uncertainty in modelled ablation was expected given the uncertainties involved in parameterising incoming radiation fluxes and albedo. A positive bias in modelled ablation rates was exhibited, though the 1 : 1 line is still well within the model uncertainty (\(2\sigma\)). This bias was likely an artefact of the limited value of the cloud extinction co-efficient (\(k\)), which produced a positive bias in ensemble mean SW\(\downarrow\) as compared to the reference run (not shown). However, this bias was of less concern as the remaining analysis used the reference run and not the ensemble mean from the Monte Carlo runs to explore cloud effects on SMB and \(\Delta\text{SMB}\). That the temporal variability of SMB was effectively captured by SEBpr gives us confidence that this configuration captures the same atmospheric controls on SMB as SEBmr and as such provides a reliable and useful tool for sensitivity analysis.
3.2 Variation of SBL climate with cloudiness

The seasonal variation of surface climate in both clear-sky and overcast conditions during the measurement period is shown in Fig. 5a and b. Air temperature ($T_a$) exhibited a clear but relatively small ($\sim 8$ °C) seasonal cycle and was only slightly lower in overcast conditions compared to clear-sky conditions (Table 5). Vapour pressure ($e_a$) was significantly higher in overcast conditions, due to the similar $T_a$ but markedly higher RH. Consequently in overcast conditions, mean $e_a$ was above the saturated vapour pressure of a melting snow/ice surface (6.11 hPa) during December through April, while in clear-sky conditions mean $e_a$ only reached this condition during February. Average $T_s$ exhibited pronounced differences, being significantly higher in overcast conditions during every month. Average wind speed ($U$) was somewhat higher (0.1 to 0.7 m s$^{-1}$) in overcast conditions during most of the ablation season, while only small or non-significant differences with cloudiness were noted in other seasons (Table 3). Thus, the main changes in surface climate observed during cloudy periods were an increase in $e_a$, which, despite slightly lower $T_a$, were associated with a large increase in $T_s$.

3.3 Variation of SEB and melt with cloudiness

Monthly average SEB terms diagnosed using SEBmr showed marked variation with cloudiness and season during the measurement period (Fig. 5c and d). Clear-sky conditions were characterised by large and opposing fluxes. SWnet dominated the seasonal cycle, provided the largest source of energy during the summer months and peaked after the summer solstice in response to decreased albedo associated with the transition from a snow to ice surface in early January. LWnet remained a large sink throughout the year, creating strongly negative Rnet during the winter months (JJA) that drove cooling of the glacier surface. Low $T_s$ in clear-sky conditions allowed QS to remain directed towards the surface throughout the year. QS was of a similar magnitude to LWnet and peaked during the winter months in response to an increase in both $U$ and the surface–air temperature gradient (Fig. 5a and b). QL was much smaller in...
magnitude than QS and of a generally negative sign, indicating that during clear-skies, sublimation or evaporation dominated over deposition or condensation. QR was absent and positive QC indicated that nocturnal cooling of the surface and subsurface was occurring. QM in excess of 20 W m\(^{-2}\) (equivalent to 5 mm w.e. day\(^{-1}\)) was present for a 7 month period between October and April (inclusive). In general the seasonal cycle of QM followed that of SWnet, but was modulated by variations in QL and QS.

In contrast to clear-skies, energy terms in overcast conditions were smaller in magnitude and directed towards the surface on average (Fig. 5d). SWnet was still the largest source of energy to the surface. LWnet was positive through most of the year, due to the enhancement of LW\(_\downarrow\) by low cloud cover and the \(T_s\) being limited to 0 °C. Consequently, Rnet was positive throughout the year and larger than in clear-sky conditions from March to November (inclusive). QS and QL were nearly equal and both directed towards the surface, together producing a similar source of energy as Rnet. A distinct seasonal cycle in QS and QL was driven by the strong seasonal variation in surface–air temperature and moisture gradients in overcast conditions (Fig. 5b). QR made a small contribution to QM during the summer and QC was negligible. The net result was that despite the moderate magnitude of individual energy fluxes in overcast conditions, mean QM was similar to that in clear-sky conditions during most months and exceeded that in clear-sky conditions from February through May.

The similarity of mean QM in clear-sky and overcast conditions was due to a large extent to the fraction of time the surface was melting in each condition (Fig. 6). In clear-sky conditions, melt occurred for a much smaller fraction of time, reaching a maximum of 58 % during December, while in overcast conditions, melt occurred between 70 and 95 % of the time over the 7 months from October through April (inclusive). Day length during this period varied between 11.5 and 15.5 h, implying that nocturnal melt in overcast conditions was a significant feature during these months. While clear-sky and overcast conditions were experienced 36 and 45 % of the measurement period (Conway et al., 2014), they accounted for 30 and 50 % of total melt, respectively, as overcast conditions accounted for a much higher proportion of all melting periods.
When all melt periods were considered together (42% of measurement period), SWnet made the largest contribution to QM, with QS and QL together contributing a little over one third and QR providing a non-negligible fraction (Table 6). On average LWnet and QC were energy sinks during melting periods, though as noted earlier they diverged strongly with cloudiness. Considering the average SEB terms in all periods, a shift towards QS at the expense of Rnet was observed, due to the inclusion of non-melting clear-sky periods, where negative LWnet was largely balanced by QS.

3.4 Sensitivity of SMB to changes in surface climate

Model runs with SEBpr over the sensitivity period (see Sect. 2.4) revealed a large ∆SMB to $T_a$ (Table 7). The response to a ±1 K change in $T_a$ was much larger than that of a 20% change in $P_{\text{scaled}}$ and this large sensitivity is explored further in the following section. Increased RH induced a small mass loss, due to increased LW↓ and QL. Similarly, a mass loss of 0.79 mwe.a$^{-1}$ occurred for a 1 m$s^{-1}$ increase in $U$, due to an increased contribution of turbulent heat fluxes to melt. The ∆SMB to terms controlling SWnet is surprisingly high, though no study has assessed this in the Southern Alps to date. A change in $\alpha$ of ±0.1 induced over half the SMB response of $T_a$ ± 1 K (Table 7), while a 6% decrease in SW$_{\text{TOA}}$ (the approximate change in the solar constant during the last 10 000 years) only resulted in a modest change in SMB.

Variations in the cloud extinction coefficient ($k$), within the uncertainty range of the radiation scheme optimisation (Conway et al., 2014), induced large mass changes, emphasising the fact that SWnet still makes the largest contribution to melt during overcast conditions (Table 6).

To examine how the large ∆SMB to $T_a$ was expressed, a breakdown of SMB terms was constructed for the +1 and −1 K perturbation runs (Table 8). A change in snowfall accounted for 21% of ∆SMB, while a small change in refreezing (2%) and a dominant change in melt accounted for the remainder (77%). Changes in deposition and sublimation were negligible. The large change in accumulation with $T_a$ emphasises the temperate nature of the glacier SBL in the vicinity of AWS$_{\text{glacier}}$, where mean $T_a$ dur-
ing precipitation is 1 °C and snow fall can occur at any time of the year (Cullen and Conway, 2015). Indeed, despite the large ablation at AWS_{glacier} over the measurement period (> 9 m.w.e.), a decrease in $T_a$ of 1.3 K was sufficient to produce a net zero SMB (not shown).

The change in melt between $T_a$ perturbation runs can be attributed to changes in QM that are direct, i.e. a change in the magnitude of LW↓, QS, QL, and QR; and indirect, i.e. a change in SWnet driven by snowdepth–α feedbacks. Table 9 shows mean SEB components for each $T_a$ perturbation run. The most striking feature is that while a 100 % increase in melt occurred between −1 and +1 K runs (Table 8), there was only a 40 % increase in QM during melt (Table 9, A and B final column). The majority of the increased melt was due to a large increase in the fraction of time melt occurred, from 34 to 48 % of all periods. Thus, a better indication of the contribution of each SEB term to $\Delta$SMB can be found by examining the change in SEB terms between runs for the melting periods in the +1 K run, (Table 9, E). SWnet showed the largest contribution to the increase in QM between these scenarios, and considering that melt made up 77 % of the total $\Delta$SMB, it follows that SWnet accounted for 36 % of the $\Delta$SMB. In the same way, QS (15 %) and QL (14 %) together accounted for less than a third of the $\Delta$SMB. Adding LWnet (9 %) and QR (2 %) to the contribution from the turbulent heat fluxes, direct changes in QM accounted for under half of the $\Delta$SMB, with changes in snowfall and indirect changes in QM dominating. Given the covariance of cloudiness and SEB terms shown in Sect. 3.3 and the obvious link between cloudiness and precipitation, further examination of the interplay between cloudiness and $\Delta$SMB is made in the following section.

3.5 Impact of cloud on SMB sensitivity

To begin to describe the influence of cloud cover on the $\Delta$SMB to $T_a$, the amount of melt that occurred under clear-sky, partial cloud and overcast conditions was calculated for each $T_a$ perturbation run (Fig. 7). In absolute terms overcast periods showed the largest change in melt between $T_a$ perturbation runs, accounting for 50 % of the $\Delta$SMB to $T_a$. 
Clear-sky and partial cloud conditions showed more modest changes in melt, accounting for 29 and 21% of the ΔSMB, respectively. By calculating the mean ΔSMB for clear-sky and overcast conditions in each month, a distinct seasonal cycle as well as a clear dependence on cloudiness emerged (Fig. 8). In general, the ΔSMB was greatly reduced during winter months, as $T_a$ was well below $T_{r/s}$ at AWS$_{glacier}$ and ablation was minimal. On average, overcast conditions almost always produced higher ΔSMB than clear-skies, especially during spring and autumn. A peak in ΔSMB observed during October was associated with $T_a$ around $T_{r/s}$ and a higher fraction of marginal melt conditions. ΔSMB in clear-sky conditions showed a long period of minimal ΔSMB from May through October (inclusive). January and February, however, show a large ΔSMB in clear-sky conditions, as the magnitude of SWnet was greatly influenced by the timing of the transition to an ice surface and summer snowfall. In order to isolate the snowdepth–α feedback, further runs of SEBpr were made for $-1$ and $+1$ K. By using measured $\alpha$ and perturbing $T_{r/s}$ with $T_a$, accumulation and SWnet remained consistent between these runs and the resulting ΔSMB (direct) was due to direct changes in QM only (Fig. 8., dashed lines). The divergence of full and direct ΔSMB in clear-skies conditions confirmed that changes in melt due to snowdepth–α feedbacks dominated clear-sky ΔSMB, especially in the summer. In overcast conditions, the direct ΔSMB is somewhat less than the full SMB, as periods with altered snowfall are removed. Still, the direct ΔSMB remained approximately twice as large as that in clear-sky conditions through each month. Thus, it is evident that cloudy conditions have a much stronger influence on ΔSMB to $T_a$ than clear-sky conditions, with an increased ΔSMB in cloudy conditions being due to changes in both snowfall and melt, and being strongest in the spring and autumn seasons.
4 Discussion

4.1 Cloud impacts on SBL and SEB

The strong divergence of SEB with cloud condition seen in these results is driven in large part by changes in $e_a$, rather than changes in $T_a$. The increase in $e_a$ in overcast conditions is enabled by the poor association of $T_a$ and cloud cover, in addition to the obvious covariance between RH and cloudiness. That $T_a$ is not markedly decreased in overcast conditions differs from similar studies in the European Alps (e.g. Pellicciotti et al., 2005) and Norway (Giesen et al., 2008), and is indicative of the maritime setting where airmass properties, rather than a positive association between summertime insolation and air temperature (Sicart et al., 2008), are the primary control on SBL variations (Cullen and Conway, 2015). The availability of moist and relatively warm air masses to the glacier surface also creates positive LWnet in overcast conditions, which along with increases in QL, allows for steady melt through much greater periods of time. Consequently, average daily melt rates are similar in clear-sky and overcast conditions, again in contrast with studies in the European Alps that show increased melt in clear-sky conditions (Pellicciotti et al., 2005). Glaciers in Norway (Giesen et al., 2008) show higher total melt during overcast conditions due to higher $U$ that increase turbulent heat fluxes during frequent cloud cover. While increased $U$ and turbulent heat fluxes are observed for the largest melt events on Brewster Glacier (Gillett and Cullen, 2011), mean $U$ was not well differentiated by cloudiness over the measurement period, leaving $T_a$ and $e_a$ as the primary controls of mean QS and QL, respectively.

While LWnet was substantially increased during overcast periods, a “radiation paradox” (Ambach, 1974) does not occur during most of the melt season at Brewster Glacier, due to high $SW_{TOA}$, large cloud extinction coefficients and a smaller difference in sky emissivity in clear-sky and overcast conditions in this mid-latitude environment. In contrast, maritime sites on the melting margin of the Greenland ice sheet show clouds act to increase $R_{net}$ throughout the melt season at a range of elevations (van den Broeke et al., 2008a). At the lowest site where the surface is melting over 80% of
the summer period, the presence of a strong “radiation paradox” implies that melt rates are higher during overcast conditions, which is supported by the absence of increased summer melt during more frequent clear-sky conditions (van den Broeke et al., 2011). The lack of a “radiation paradox” during the summer months at Brewster Glacier emphasises the role of airmass properties that are advected from the surrounding ocean areas in maintaining $T_a$ and enabling enhanced LWnet and QL during overcast periods. In the same way, during the transition periods, especially in the autumn, increased melt rates were enabled by a “radiation paradox”.

4.2 Cloud impacts on SMB sensitivity

The increased sensitivity of SMB to $T_a$ in overcast conditions may help explain some of the high sensitivity of SMB in the Southern Alps. Importantly, average melt is not reduced in overcast conditions and cloud cover is frequent in the Southern Alps. Therefore, a large fraction of melt occurs in overcast conditions which are more sensitive to changes in $T_a$. In conjunction with increased $e_a$, clouds extend melt into periods of marginal melt that are more sensitive to changes in $T_a$ as well as being strongly associated with frequent precipitation around $T_{r/s}$. Indeed, roughly half of the $T_a$ sensitivity is due to accumulation–albedo feedbacks, in line with previous work in the Southern Alps (Oerlemans, 1997), emphasising the turbulent heat fluxes play a secondary role, despite the assertions of recent paleo-climatic research (Putnam et al., 2012). In addition, the largest melt events – which constitute a large fraction of melt over a season (Gillett and Cullen, 2011) – are associated with overcast conditions and contribute to proportionally larger changes in melt. Thus, airmass variability, in particular air temperature associated with high water vapour content, appears to be the primary control on melt during the summer ablation season.

Aside from their role in the $\Delta$SMB to $T_a$, the contribution of turbulent heat fluxes to melt appears to have been overstated in a number of studies, at the expense of $R_{net}$. In fact, the contribution of $R_{net}$ to ablation in the present study is similar to that found over mixed snow/ice ablation surfaces in Norway (68%; Giesen et al., 2008)
and coastal Greenland (~ 70% (S6); van den Broeke et al., 2008b), and similar to that found for a neve area in New Zealand (Kelliher et al., 1996). There are a number of possible reasons for the deviation of the current study from previously reported values for glacier surfaces in the Southern Alps (e.g. Marcus et al., 1985; Hay and Fitzharris, 1988; Ishikawa et al., 1992; Anderson et al., 2010). Firstly, in earlier studies simplifications were usually made in the calculation of the turbulent heat fluxes, including the assumption that the surface is always melting. Secondly, average SEB terms were traditionally reported for the entire study period, rather than only those during periods of melt. Table 6 clearly shows full-period average SEB terms are biased towards turbulent heat fluxes, as non-melting nocturnal and winter periods where enhanced QS balances negative LWnet are included. Lastly, a number of the studies have been conducted in low elevation areas where turbulent heat fluxes are increased, despite these areas being atypical in the Southern Alps (mean elevation of glacier termini > 1500 m a.s.l.; Hoelzle et al., 2007).

4.3 Implications for modelling glacier–climate interactions

While the present study does not make an assessment of glacier wide ∆SMB and therefore is somewhat limited in discussing atmospheric controls on glacier fluctuations, it shows that the response of glacier melt to changes in $T_a$ can be altered by clouds. This has two implications for our understanding of glacier climate interactions. Firstly, efforts to characterise glacier–climate connections need to consider the effects of changing atmospheric moisture on melt rate as well as accumulation. New avenues to model glacier melt with enhanced temperature index models (TIM) or other empirical descriptions of the temperature dependant fluxes (e.g. Giesen and Oerlemans, 2012) need to consider the variance of atmospheric moisture with respect to melt. This is both due to the strong increase in LW↓ by clouds, but also the association with increased positive QL in moist environments. This may be important for other maritime areas, as well as the Southern Alps where TIM’s have already been shown to break down in large melt events (Cutler and Fitzharris, 2005; Gillett and Cullen, 2011).
Secondly, it follows that a change in the frequency of cloud cover or synoptic regime may enhance/dampen the SMB response to $T_a$. For example, a decrease in $\Delta$SMB from west to east across the Southern Alps is likely, in association with the strong gradient of precipitation and cloudiness (Uddstrom et al., 2001). It is enticing to reduce the relationship between glacier mass balance and climate to the main causal mechanisms (i.e. temperature/precipitation paradigm). However, there is also the possibility that changes in atmospheric circulation coincident with changes in state variables in the past (i.e. during the last glacial maximum; Drost et al., 2007; Ackerley et al., 2011) may alter empirical relationships (i.e. TIM's) informed during the present climate, altering the climate signals derived from glacier fluctuations. For the Southern Alps, the most compelling analysis of the controls on SMB points to changes in the regional circulation patterns (Fitzharris et al., 2007), which are in turn associated with strong changes in both airmass properties and cloudiness (Hay and Fitzharris, 1988). Thus, it is likely that average relationships between melt and air temperature may indeed be changed if a shift to drier or wetter conditions is experienced.

The high fraction of melt due to SWnet and large contribution of snowfall–albedo feedbacks to $\Delta$SMB also implies that local or regional influences on albedo may have an important role in modifying melt rate as seen in other areas (Oerlemans et al., 2009). Indeed the LGM period shows higher rates of glacial loess deposition in New Zealand (Eden and Hammond, 2003), thus the role of terrigenous dust in modifying glacier ablation rates during the onset of glacier retreat (e.g. Peltier and Marshall, 1995) is a topic that should be explored further in the context of the Southern Alps.

5 Conclusions

We have presented here a validated timeseries of SEB/SMB for a glacier surface in the Southern Alps of New Zealand during 2 annual cycles. High quality radiation data allowed a careful evaluation of the magnitude of SEB terms, as well as the selection of clear-sky and overcast conditions. An analysis of SBL climate and SEB showed a fun-
damental change in SEB with cloudiness that was driven by an increase in effective sky emissivity and vapour pressure at the glacier surface. The only slightly diminished $T_a$ during overcast periods created positive LWnet and also allowed both QS and QL to remain large and directed toward the surface. This created a strong increase in the fraction of time the surface was melting in overcast conditions, which led to a similar average melt rate in clear-sky and overcast conditions. Given the frequent cloud cover at the site, cloudy periods accounted for a majority of the melt observed, especially during autumn when SWnet inputs were lower.

A parameterisation of radiation components allowed the sensitivity of SMB to independent changes in SBL climate and shortwave radiation components to be assessed. The large sensitivity of SMB to $T_a$ was expressed primarily through changes in snowfall and the associated positive $\alpha$ feedback. The remainder of this sensitivity was due to changes in the fraction of time the surface was melting and changes in the magnitude of QS, QL, LWnet and QR, in that order. The sensitivity of SMB to $T_a$ diverged strongly when partitioned into clear-sky and overcast periods, with enhanced sensitivity during overcast periods due to both their covariance with precipitation and their ability to produce melt over large fractions of time. Increased sensitivity during overcast periods may explain some of the high sensitivity of SMB in the Southern Alps, and raises the possibility that the response of SMB to $T_a$ in the past or future may be altered by changing synoptic patterns that are strongly associated with cloud cover. Thus, it highlights the need to include the effect of atmospheric moisture (vapour, cloud and precipitation) on both melt and accumulation processes when modelling glacier–climate interactions.

Acknowledgements. Funding from a University of Otago Research Grant (ORG10-10793101RAT) supported N. Cullen’s contribution to this research. The research also benefited from the financial support of the National Institute of Water and Atmospheric Research, New Zealand (Climate Present and Past CLC01202), and collaboration with A. Mackintosh and B. Anderson, Victoria University of Wellington. The Department of Conservation supported this research under concession OT-32299-OTH. We thank T. Mölg for providing model code and for many helpful discussions on model development, as well as P. Sirguey and W. Colgan for help-
ful discussions on the Monte Carlo approach. D. Howarth and N. McDonald provided careful technical support for the field measurements.

References


Cloud effects on the surface energy and mass balance of Brewster Glacier

J. P. Conway and N. J. Cullen

Abstract

Introduction

Conclusions

References

Tables

Figures
Cloud effects on the surface energy and mass balance of Brewster Glacier

J. P. Conway and N. J. Cullen


Cloud effects on the surface energy and mass balance of Brewster Glacier

J. P. Conway and N. J. Cullen

References


Table 1. Variables measured, sensor specifications and mean annual values at AWS\textsubscript{glacier}.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Instrument</th>
<th>Accuracy</th>
<th>Mean annual value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air temperature ($T_a$)</td>
<td>Vaisala HMP 45AC</td>
<td>0.3°C</td>
<td>1.2°C</td>
</tr>
<tr>
<td>Relative humidity (RH)</td>
<td>Vaisala HMP 45AC</td>
<td>3%</td>
<td>78%</td>
</tr>
<tr>
<td>Wind speed ($U$)</td>
<td>RM Young 05103</td>
<td>0.3 m\textsuperscript{s}\textsuperscript{-1}</td>
<td>3.3 m\textsuperscript{s}\textsuperscript{-1}</td>
</tr>
<tr>
<td>Atmospheric pressure ($p$)</td>
<td>Vaisala PTB110</td>
<td>0.5 hPa</td>
<td>819 hPa</td>
</tr>
<tr>
<td>Incoming shortwave radiation ($SW\downarrow_{\text{meas}}$)</td>
<td>Kipp and Zonen CNR4</td>
<td>5%\textsuperscript{a}</td>
<td>140 W m\textsuperscript{-2}</td>
</tr>
<tr>
<td>Outgoing shortwave radiation ($SW\uparrow$)</td>
<td>Kipp and Zonen CNR4</td>
<td>5%\textsuperscript{a}</td>
<td>93 W m\textsuperscript{-2}</td>
</tr>
<tr>
<td>Incoming longwave radiation ($LW\downarrow_{\text{meas}}$)</td>
<td>Kipp and Zonen CNR4</td>
<td>5%\textsuperscript{a}</td>
<td>278 W m\textsuperscript{-2}</td>
</tr>
<tr>
<td>Surface temperature ($T_s$)</td>
<td>Kipp and Zonen CNR4</td>
<td>1°C\textsuperscript{b}</td>
<td>−2.7°C</td>
</tr>
<tr>
<td>Precipitation ($P_{\text{scaled}}$)\textsuperscript{c}</td>
<td>TB4 + Scaled\textsuperscript{c}</td>
<td>25%\textsuperscript{c}</td>
<td>6125 mm\textsuperscript{c}</td>
</tr>
<tr>
<td>Surface and sensor height</td>
<td>SR50a</td>
<td>±1 cm</td>
<td>n/a</td>
</tr>
</tbody>
</table>

\textsuperscript{a} Uncertainty is estimated to be less than the manufacturer’s specifications as noted in van den Broeke et al. (2004) and Blonquist et al. (2009).

\textsuperscript{b} Based on a 5 W m\textsuperscript{-2} uncertainty in LW\textsuperscript{↑}.

\textsuperscript{c} From AWS\textsubscript{lake} during snow free period only. $P_{\text{scaled}}$ is based on scaled relationship between AWS\textsubscript{lake} and lowland station (Cullen and Conway, 2015). Uncertainty is estimated from fit of scaled relationship.
Table 2. Configuration of SEBmr and SEBpr, showing input data and references used in the calculation of radiation terms in each configuration.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Model version</th>
<th>Reference and/or input data</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$</td>
<td>SEBmr</td>
<td>Accumulated albedo (van den Broeke et al., 2004)</td>
</tr>
<tr>
<td></td>
<td>SEBpr</td>
<td>Oerlemans and Knap (1998) ($P_{\text{scaled}}, T_a$)</td>
</tr>
<tr>
<td>$SW_{\downarrow}$</td>
<td>SEBmr</td>
<td>$SW_{\downarrow,\text{surface}}$</td>
</tr>
<tr>
<td></td>
<td>SEBpr</td>
<td>Conway et al. (2014) ($N_e, T_a, \text{RH}$)</td>
</tr>
<tr>
<td>$LW_{\downarrow}$</td>
<td>SEBmr</td>
<td>$LW_{\downarrow,\text{meas}}$</td>
</tr>
<tr>
<td></td>
<td>SEBpr</td>
<td>Conway et al. (2014) ($N_e, T_a, \text{RH}$)</td>
</tr>
</tbody>
</table>
Table 3. Input parameter uncertainty introduced in Monte Carlo simulations of SMB uncertainty.

<table>
<thead>
<tr>
<th>Input parameter</th>
<th>Value(s)</th>
<th>Systematic [random] error</th>
<th>Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Roughness length for momentum ($z_0v$)(^a)</td>
<td>3.6 × 10(^{-3}) m</td>
<td>$z_0v \times 10^\text{NORMRND}(0,0.274)$</td>
<td>SEBmr, pr</td>
</tr>
<tr>
<td>Rain/snow threshold ($T_{rs}$)(^b)</td>
<td>1.0 °C</td>
<td>0.3 [0.5]</td>
<td>SEBmr, pr</td>
</tr>
<tr>
<td>Albedo of surface ($\alpha_{\text{snow}}, \alpha_{\text{firm}}, \alpha_{\text{ice}}$)(^b)</td>
<td>0.95 ($\alpha_{\text{snow}}$) 0.65 ($\alpha_{\text{firm}}$) 0.42 ($\alpha_{\text{ice}}$)</td>
<td>0.05</td>
<td>SEBpr</td>
</tr>
<tr>
<td>Constant for cloud extinction coefficient(^c)</td>
<td>0.1715</td>
<td>0.0048 [0.0048]</td>
<td>SEBpr</td>
</tr>
<tr>
<td>Multiplier for cloud extinction coefficient(^c)</td>
<td>0.07182 hPa(^{-1})</td>
<td>0.0324 [0.0324]</td>
<td>SEBpr</td>
</tr>
<tr>
<td>Albedo of surrounding terrain(^d)</td>
<td>0.45</td>
<td>0.1</td>
<td>SEBpr</td>
</tr>
<tr>
<td>Clear-sky emissivity constant(^e)</td>
<td>0.456 Pa(^{-1}) K</td>
<td>0.0204 [0.0204]</td>
<td>SEBpr</td>
</tr>
</tbody>
</table>

\(^a\) SD of $z_0v$ (Conway and Cullen, 2013). NORMRND is a MATLAB function that selects a random number from a normal distribution with mean of 0 and SD of 0.274.

\(^b\) Macguth et al. (2008).

\(^c\) 95% confidence interval of optimised coefficients (Conway et al., 2014). Limited to 0.95.

\(^d\) Assumed; no random errors as terrain albedo will not vary at this timescale (30 min).

\(^e\) RMSD of clear-sky values (Conway et al., 2014).
Table 4. Observed and modelled SMB (m.w.e.) for selected periods between stake measurements in ablation (Abl) and accumulation (Acc) seasons. Figure 2b shows the length of each period.

<table>
<thead>
<tr>
<th>Period</th>
<th>Observed</th>
<th>SEBmr</th>
<th>SEBpr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abl 1 snow</td>
<td>−1.74</td>
<td>−1.78</td>
<td>−1.67</td>
</tr>
<tr>
<td>Abl 1 ice</td>
<td>−3.35</td>
<td>−2.92</td>
<td>−3.28</td>
</tr>
<tr>
<td>Acc1</td>
<td>1.52</td>
<td>1.40</td>
<td>1.46</td>
</tr>
<tr>
<td>Abl 2 snow</td>
<td>−1.51</td>
<td>−1.78</td>
<td>−1.48</td>
</tr>
<tr>
<td>Abl 2 ice</td>
<td>−1.94</td>
<td>−1.87</td>
<td>−1.66</td>
</tr>
</tbody>
</table>
Cloud effects on the surface energy and mass balance of Brewster Glacier

J. P. Conway and N. J. Cullen

Table 5. Mean differences in surface climate between clear-sky and overcast conditions. Positive values indicate an increase in overcast conditions.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Jan</th>
<th>Feb</th>
<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
<th>Aug</th>
<th>Sep</th>
<th>Oct</th>
<th>Nov</th>
<th>Dec</th>
<th>Annual</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_a$ (°C)</td>
<td>-1.3</td>
<td>-0.2</td>
<td>-0.8</td>
<td>-3</td>
<td>-0.9</td>
<td>-0.5</td>
<td>-1.3</td>
<td>-1.2</td>
<td>-0.7</td>
<td>-0.1</td>
<td>-1.4</td>
<td>-1.2</td>
<td>-1.1</td>
</tr>
<tr>
<td>RH (%)</td>
<td>35</td>
<td>25</td>
<td>35</td>
<td>53</td>
<td>44</td>
<td>39</td>
<td>52</td>
<td>37</td>
<td>45</td>
<td>33</td>
<td>34</td>
<td>37</td>
<td>39</td>
</tr>
<tr>
<td>$e_a$ (hPa)</td>
<td>2.5</td>
<td>2.2</td>
<td>2.5</td>
<td>3.1</td>
<td>2.6</td>
<td>2.2</td>
<td>2.6</td>
<td>1.7</td>
<td>2.3</td>
<td>2.1</td>
<td>2</td>
<td>2.7</td>
<td>2.4</td>
</tr>
<tr>
<td>$U$ (m s$^{-1}$)</td>
<td>0.6</td>
<td>0.1</td>
<td>0.6</td>
<td>-0.2</td>
<td>-0.1</td>
<td>0</td>
<td>0</td>
<td>-0.2</td>
<td>-0.1</td>
<td>-0.1</td>
<td>0.2</td>
<td>0.7</td>
<td>0.1</td>
</tr>
<tr>
<td>$T_s$ (°C)</td>
<td>0.5</td>
<td>0.3</td>
<td>0.9</td>
<td>1.5</td>
<td>4.7</td>
<td>5.4</td>
<td>6.3</td>
<td>5.2</td>
<td>5.8</td>
<td>2.6</td>
<td>1.5</td>
<td>0.6</td>
<td>2.9</td>
</tr>
<tr>
<td>$p$ (hPa)</td>
<td>-7</td>
<td>-4</td>
<td>-8</td>
<td>-6</td>
<td>-10</td>
<td>-7</td>
<td>-12</td>
<td>-4</td>
<td>-9</td>
<td>-3</td>
<td>-7</td>
<td>-8</td>
<td>-7</td>
</tr>
</tbody>
</table>

Bold face indicates monthly differences are significant at the 95 % level using a two sided $t$ test assuming unequal variances. Temperature and wind speed are normalised to 2 m values.
Table 6. Average surface energy fluxes (W m\(^{-2}\)) for melting periods in clear-sky and overcast conditions, all melting periods, and all periods during the measurement period. Bracketed bolds show the proportion of QM for each condition.

<table>
<thead>
<tr>
<th></th>
<th>SWnet</th>
<th>LWnet</th>
<th>Rnet</th>
<th>QS</th>
<th>QL</th>
<th>QR</th>
<th>QC</th>
<th>QM</th>
</tr>
</thead>
<tbody>
<tr>
<td>Melting + clear-sky periods</td>
<td>240 (121)</td>
<td>-67 (−34)</td>
<td>173 (87)</td>
<td>39 (20)</td>
<td>-7 (−3)</td>
<td>0 (0)</td>
<td>-6 (−3)</td>
<td>199</td>
</tr>
<tr>
<td>Melting + overcast periods</td>
<td>36 (33)</td>
<td>15 (14)</td>
<td>51 (46)</td>
<td>30 (27)</td>
<td>24 (22)</td>
<td>7 (7)</td>
<td>-2 (−2)</td>
<td>110</td>
</tr>
<tr>
<td>Melting periods</td>
<td>96 (70)</td>
<td>-8 (−6)</td>
<td>88 (65)</td>
<td>32 (24)</td>
<td>15 (11)</td>
<td>5 (3)</td>
<td>-3 (−2)</td>
<td>136</td>
</tr>
<tr>
<td>All periods</td>
<td>49 (83)</td>
<td>-27 (−46)</td>
<td>22 (37)</td>
<td>31 (53)</td>
<td>2 (3)</td>
<td>2 (4)</td>
<td>2 (3)</td>
<td>58</td>
</tr>
</tbody>
</table>

Melting conditions are selected as periods where QM > 0 in SEBmr.
Table 7. ΔSMB (mm w.e. a⁻¹) to changes in surface climate and shortwave radiation terms. Values are averages of positive and negative perturbation runs of SEBpr over the sensitivity period, while the sign of ΔSMB is shown for an increased in each input variable or parameter.

<table>
<thead>
<tr>
<th>Variable and perturbation</th>
<th>ΔSMB</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_a \pm 1$ K</td>
<td>−2065</td>
</tr>
<tr>
<td>$P_{\text{scaled}} \pm 20%$</td>
<td>+770</td>
</tr>
<tr>
<td>RH $\pm 10%$</td>
<td>−380</td>
</tr>
<tr>
<td>$U \pm 1$ m s⁻¹</td>
<td>−790</td>
</tr>
<tr>
<td>$\alpha \pm 0.1$</td>
<td>+1220</td>
</tr>
<tr>
<td>Solar constant $-6%$</td>
<td>−260</td>
</tr>
<tr>
<td>$k \pm 0.17$</td>
<td>−740</td>
</tr>
</tbody>
</table>
Table 8. Sum of SMB terms for selected runs of SEBpr over the two year sensitivity period. All units are in mm w.e., except for Δ which is in mm w.e. K⁻¹ a⁻¹.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>SMB</th>
<th>Snowfall</th>
<th>Melt</th>
<th>Sublimation</th>
<th>Deposition</th>
<th>Refreezing</th>
</tr>
</thead>
<tbody>
<tr>
<td>+1 K</td>
<td>-9181</td>
<td>3900</td>
<td>13064</td>
<td>32</td>
<td>134</td>
<td>85</td>
</tr>
<tr>
<td>-1 K</td>
<td>-920</td>
<td>5670</td>
<td>6992</td>
<td>38</td>
<td>135</td>
<td>198</td>
</tr>
<tr>
<td>Δ (mm w.e. K⁻¹ a⁻¹)</td>
<td>-2065</td>
<td>443</td>
<td>-1593</td>
<td>2</td>
<td>0</td>
<td>28</td>
</tr>
</tbody>
</table>
Table 9. Mean SEB terms during melting periods in the +1 K (A) and −1 K (B) perturbation runs of SEBpr. Also shown are mean SEB terms in the −1 K perturbation run, for the same periods as A, i.e. melting periods in the +1 K perturbation run (C), and the increases between each scenario (D, E). The contribution of each flux to QM, or the increase in QM, is given in bracketed bolds.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>SWnet</th>
<th>LWnet</th>
<th>Rnet</th>
<th>QS</th>
<th>QL</th>
<th>QR</th>
<th>QC</th>
<th>QM</th>
</tr>
</thead>
<tbody>
<tr>
<td>A: +1 K melting periods</td>
<td>89 (62)</td>
<td>−4 (−3)</td>
<td>85 (59)</td>
<td>37 (26)</td>
<td>19 (14)</td>
<td>5 (3)</td>
<td>−2 (−1)</td>
<td>144</td>
</tr>
<tr>
<td>B: −1 K melting periods</td>
<td>70 (68)</td>
<td>−9 (−8)</td>
<td>61 (60)</td>
<td>28 (27)</td>
<td>11 (11)</td>
<td>4 (4)</td>
<td>−2 (−2)</td>
<td>103</td>
</tr>
<tr>
<td>C: −1 K for same periods as A</td>
<td>56 (76)</td>
<td>−13 (−17)</td>
<td>43 (59)</td>
<td>23 (32)</td>
<td>7 (9)</td>
<td>3 (4)</td>
<td>−2 (−2)</td>
<td>74</td>
</tr>
<tr>
<td>D: increase from B to A</td>
<td>19 (46)</td>
<td>5 (11)</td>
<td>23 (57)</td>
<td>9 (22)</td>
<td>8 (20)</td>
<td>1 (2)</td>
<td>0 (0)</td>
<td>41</td>
</tr>
<tr>
<td>E: increase from C to A</td>
<td>33 (47)</td>
<td>8 (12)</td>
<td>41 (59)</td>
<td>14 (20)</td>
<td>13 (18)</td>
<td>2 (3)</td>
<td>0 (0)</td>
<td>70</td>
</tr>
</tbody>
</table>
Figure 1. Map of Brewster Glacier showing AWS locations and surrounding topography. Contour lines are at 100 m intervals. Long-term mass balance network (MB stakes) shown as filled circles. The glacier margin shown is based on a 1997 GPS survey (Willis et al., 2009). The ridgeline to the southeast of the glacier is the main divide of the Southern Alps. The inset map shows the location of Brewster Glacier within New Zealand.
Cloud effects on the surface energy and mass balance of Brewster Glacier

J. P. Conway and N. J. Cullen

Figure 2. (a) Daily average albedo observed at AWS\textsubscript{glacier} (red) during the measurement period and modelled in SEBpr (blue) using the expressions of Oerlemans and Knap (1998), with optimised coefficients. (b) Accumulated SMB during the measurement period as modelled by the reference runs of SEBmr and SEBpr. The points give observed mass balance from periodic stake and snow pit measurements. The SMB for selected ablation and accumulation periods (shown as Abl1 snow etc.) are given in Table 4. The shaded envelope shows ±1 SD from the mean of SEBmr, calculated using Monte Carlo simulations (see Sect. 2.4 for details).
Figure 3. Observed vs. modelled surface temperature for (a) SEBmr and (b) SEBpr runs. Red dots are 30 min averages, while black dots are monthly averages.
Figure 4. Observed vs. modelled mass balance for (a, b) SEBmr and (c, d) SEBpr over 1 day and 10 day periods. Error bars show ±2σ from the ensemble mean values.
Figure 5. Monthly mean surface climate (a, b) and surface energy fluxes (c, d) at AWS$_{\text{glacier}}$ in (a, c) clear-sky and (b, d) overcast conditions. Partial cloud conditions are a graduation between the two extremes and are not shown for brevity. Surface climate variables include air and surface temperature ($^\circ$C), wind speed ($\text{m}^2\text{s}^{-1}$), vapour pressure (hPa), and relative humidity on a scale from 0 to 10 (i.e. %/10).
Figure 6. Fraction of time surface melting occurred in clear-sky (open circles) and overcast (closed circles) conditions during each month. Melting conditions are selected as periods where QM > 0 in SEBmr.
Cloud effects on the surface energy and mass balance of Brewster Glacier

J. P. Conway and N. J. Cullen

Figure 7. Total surface melt in each cloud cover category for reference and climate perturbation scenarios.
Figure 8. The mean daily mass balance sensitivity ($\Delta$SMB) to a 1 K change in $T_a$, separated into clear-sky (green) and overcast (blue) conditions, in each month of the year. The dashed lines show $\Delta$SMB resulting from only a direct change in $Q_M$, which was derived from a further model run using measured albedo and perturbing $T_r/s$ with $T_a$. The positive values indicate mass loss for increased $T_a$. 