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Global application of a surface mass balance model using gridded climate data

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Abstract

Global applications of surface mass balance models have large uncertainties, as a result of poor climate input data and limited availability of mass balance measurements. This study addresses several possible consequences of these limitations for the modelled mass balance. This is done by applying a simple surface mass balance model that only requires air temperature and precipitation as input data, to glaciers in different regions. In contrast to other models used in global applications, this model separately calculates the contributions of net solar radiation and the temperature-dependent fluxes to the energy balance. We derive a relation for these temperature-dependent fluxes using automatic weather station (AWS) measurements from glaciers in different climates. With local, hourly input data, the model is well able to simulate the observed seasonal variations in the surface energy and mass balance at the AWS sites. Replacing the hourly local data by monthly gridded climate data removes summer snowfall and winter melt events and hence influences the modelled mass balance most on locations with a small seasonal temperature cycle. Representative values for the multiplication factor and vertical gradient of precipitation are determined by fitting modelled winter mass balance profiles to observations on 80 glaciers in different regions. For 72 of the 80 glaciers, the precipitation provided by the climate data set has to be multiplied with a factor above unity; the median factor is 2.55. The vertical precipitation gradient ranges from negative to positive values, with more positive values for maritime glaciers and a median value of 1.5 $\mathrm{mm\,a^{-1}\,m}$. With calibrated precipitation, the modelled annual mass balance gradient closely resembles the observations on the 80 glaciers, the absolute values are matched by adjusting either the incoming solar radiation, the temperature-dependent flux or the air temperature. The mass balance sensitivity to changes in temperature is particularly sensitive to the chosen calibration method, emphasizing the importance of well-calibrated model parameters. We additionally calculate the mass balance sensitivity to changes in incoming solar radiation, revealing that

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widely observed variations in irradiance can affect the mass balance by a magnitude comparable to a 1 °C change in temperature or a 10 % change in precipitation.

1 Introduction

The application of a glacier mass balance model on a global scale is a challenging exercise. Glaciers are situated in a variety of climates, from warm and wet to cold and dry and with seasonal cycles in temperature and/or humidity. Since the dominant processes in the surface energy and mass balance differ amongst these climates, a model should resolve all these processes for an accurate simulation of all glaciers. On the other hand, the detailed input data required for such simulations is simply not available. Meteorological measurements in mountainous terrain are scarce and suffer from local effects, whereas the spatial resolution of global and regional climate models is too coarse to resolve the specific weather conditions on glaciers. Furthermore, surface mass balance models can only be calibrated and validated on a limited sample of the world's glaciers, where meteorological and mass balance measurements have been done.

Despite these limitations, globally applied mass balance models are needed for producing estimates of the glacier contribution to sea-level rise in the next centuries. Due to the large uncertainty in meteorological data for mountainous regions, changes in the surface mass balance are generally computed with simple models, only requiring air temperature and precipitation as climatic input data (Raper and Braithwaite, 2006; Radić and Hock, 2011; Slangen et al., 2011). Climate data is often used at a low temporal (e.g., monthly) resolution, either limited by the temporal resolution of the input data or to keep the modelling time within feasible limits. Model calibration commonly relies on the available mass balance measurements. These have only been acquired at a small number of glaciers, with the longest series for the European Alps and Scandinavia and limited data for heavily glacierized regions like the Himalaya or Alaska (Zemp et al., 2008).

In this paper, we explore the applicability of a surface mass balance model in different climatic regions. The model only requires air temperature and precipitation as input data. The surface energy balance is separated into a contribution by net solar radiation and a contribution by the fluxes dependent on air temperature. In contrast to models that only use air temperature data to calculate ablation, the model used here includes the effect of the seasonal cycle in insolation (which is generally asynchronous to the annual temperature cycle) on surface melt. The model parameters are derived from automatic weather station records from different climatic regions.

We address several issues encountered when applying a mass balance model to regions with limited availability of meteorological measurements and mass balance data. We investigate the dependence of the results on the temporal resolution of the input data by comparing results obtained with hourly and monthly temperature data. By substituting the locally measured input data by climate data from the nearest grid point, we demonstrate the potential errors introduced by using meteorological data from outside the glacier environment. Next, available altitudinal profiles of winter mass balance are used to estimate the vertical precipitation gradient and a precipitation multiplication factor for 80 glaciers in different regions. The measured annual mass balance profile is matched by calibrating three different model parameter sets. Finally, the mass balance sensitivity to changes in air temperature, precipitation and atmospheric transmissivity is assessed for the sample of 80 glaciers. Special attention is given to the effect of the parameter calibration on the calculated sensitivities.

2 Mass balance model

2.1 Model description

The mass balance model is an adapted version of the simplified mass balance model described in Oerlemans (2001, p. 48). The annual surface mass balance (B) is given

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by

$$B = \int_{\text{year}} \left\{ P_{\text{snow}} + (1-r)\min\left(0; -\frac{Q}{\rho_{\text{w}} L_{\text{f}}}\right) \right\} \mathrm{d}t\,, \tag{1}$$

with mass gain resulting from solid precipitation P_{snow} and mass loss determined from the surface energy balance Q. Precipitation falls as snow when the air temperature is below a threshold temperature T_{snow}. Melt is assumed to occur whenever the surface energy balance is positive and part of the meltwater r is allowed to refreeze within the snowpack. The constant L_{f} is the latent heat of fusion, ρ_{w} is the water density (Table 1).

The energy available for melt at the surface is determined from a simplified representation of the surface energy balance, calculated at hourly time-steps. Because accurate humidity, cloudiness and wind speed data are often not available for glacierized areas, the model is set up in such a way that it only requires air temperature and precipitation data as input. Since net solar radiation is not an explicit function of air temperature, it is treated separately from the other fluxes in the energy balance, which are directly dependent on air temperature. Hence, the surface energy balance is divided into two terms (Oerlemans, 2010, p. 100):

$$Q = (1-\alpha)\tau S_{\text{in,TOA}} + \psi \tag{2}$$

where the first term describes the net solar radiative flux and the second term represents all other atmospheric fluxes, dependent on air temperature T_{a}.

Net solar radiation is computed by multiplying the incoming solar radiation at the top of the atmosphere ($S_{\text{in,TOA}}$) by the atmospheric transmissivity τ and subtracting the part of the incoming solar radiation reflected by the surface with albedo α. The atmospheric transmissivity represents all processes that affect solar radiation from the top of the atmosphere to the glacier surface, including attenuation by scattering and absorption in the clear-sky atmosphere, the transmission of radiation through clouds and shading by the topography. $S_{\text{in,TOA}}$ is computed from standard astronomical relations (e.g. Iqbal,

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1983), for τ we use one value without seasonal variation. For studies of individual glaciers, τ can be calculated in a more sophisticated way, but for global applications, the required input data (e.g., cloud observations, topography) is generally not available. When no snow is present, a constant ice albedo is used. After a snowfall event, α decreases exponentially from the fresh snow albedo α_{frsnow} to the firn albedo α_{firn}, controlled by the time-scale t_* (Oerlemans and Knap, 1998). For small snow depths, α is a function of both snow and ice albedo, according to the depth-scale d_*. The fresh snow albedo is lowered for snowfall events at temperatures around the melting point by making α_{frsnow} dependent on the air temperature during snowfall (Giesen and Oerlemans, 2010).

The temperature-dependent energy fluxes are represented by a function ψ derived from measurements at weather stations on 11 glaciers in different climates (Sect. A):

$$\psi = \begin{cases} \psi_{\mathrm{min}} + cT_{\mathrm{a}} & \text{for} \quad T_{\mathrm{a}} \geqslant T_{\mathrm{tip}}\,; \\ \psi_{\mathrm{min}} & \text{for} \quad T_{\mathrm{a}} < T_{\mathrm{tip}}\,. \end{cases} \tag{3}$$

Hence, for air temperatures below a threshold temperature T_{tip}, ψ has a constant value ψ_{min}. For higher temperatures, ψ increases linearly with T_{a}, the rate of increase given by c. Representative values for ψ_{min}, c and T_{tip} for the non-tropical AWS sites are given in Table 2.

When the right-hand-side of Eq. (2) is positive, meltwater is formed. In case the surface consists of ice, the water runs off immediately. If snow is present, a fraction r of the meltwater refreezes and heats the snowpack, while the remaining meltwater runs off. Following Oerlemans (1991), r depends on the subsurface layer temperature T_{sub} (in °C):

$$r = 1 - \exp\{T_{\mathrm{sub}}\}\,. \tag{4}$$

The change in T_{sub} resulting from refrozen meltwater is calculated as

$$\frac{\mathrm{d}T_{\mathrm{sub}}}{\mathrm{d}t} = \frac{rQ}{\rho_{\mathrm{ice}} c_{\mathrm{ice}} \delta z}\,, \tag{5}$$

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where ρ_{ice} is ice density, c_{ice} the specific heat capacity of ice and δz the subsurface layer depth (Table 1). At the end of the summer season (here defined as calendar days 121 (1 May) and 305 (1 November) in the Southern and Northern Hemispheres, respectively), T_{sub} is reset to the annual mean air temperature. If this temperature is higher than the melting point, T_{sub} is set to the melting point and refreezing will not occur at this location. This simplification of the refreezing process was shown to provide a good approximation of superimposed ice formation calculated with a sophisticated snowpack model (Wright et al., 2007).

3 Meteorological input data

In this section, the three different meteorological input data sets are described. The first two datasets are used to determine values for the model parameters and to examine the model performance at the AWS sites. The gridded data is additionally used in the global application of the mass balance model.

3.1 Hourly AWS data

The applicability of the albedo routine and the relation for ψ at the different AWS locations was examined using incoming solar radiation and air temperature measured by the AWS as model input. This allowed us to compare the measured and simulated seasonal cycles of net solar radiation and ψ without having to consider the effects of uncertainties in the input data. The meteorological records from Glaciar Lengua and Belcher Glacier are too short to be included in the analysis. For each of the other sites, a representative value for α_{ice} was determined from measured ice albedo (Table 2). These values correspond to the local surface characteristics and are partly determined by the location of the AWS with respect to the equilibrium line and the glacier tongue. Annual precipitation $P_{ann,AWS}$ (Table 3) was distributed equally over the year and chosen such that the modelled snow depth at the beginning of the ablation season (1

May) matched the measured accumulation. This allowed us to compare modelled and measured ablation in the main melt season.

3.2 Monthly AWS data

The effect of the temporal variability in the input data on the modelled mass balance was investigated by using mean seasonal cycles for air temperature and diurnal temperature range. These are based on monthly mean values computed from the AWS data, averaged over all years in each record. The daily temperature cycle was prescribed as a sine function, with the amplitude determined by the monthly mean daily temperature range. Incoming solar radiation was calculated from $S_{\mathrm{in,TOA}}$ and a constant value of τ, computed as the ratio of the annual sums of measured incoming solar radiation and $S_{\mathrm{in,TOA}}$ (Table 2). For precipitation, we used the same annual value as for the hourly AWS data.

3.3 Monthly gridded global climate data

To calculate surface mass balances on a global scale, we used a high-resolution (10′) data set on air temperature, diurnal temperature range and precipitation from the Climate Research Unit, University of East Anglia (CRU CL 2.0, New et al., 2002), hereafter referred to as CRU data. This dataset is based on measurements at a large number of weather stations, interpolated to a regular grid with sophisticated techniques. Data is available for all land areas outside Antarctica. We used the average monthly values for the period 1961–1990. Incoming solar radiation was computed in the same way as for the monthly AWS data.

For the simulations at the AWS locations, we extracted the CRU seasonal cycles of air temperature, daily temperature range and precipitation from the nearest gridpoint. The gridpoint elevations differ by −157 m (S5 Greenland) to +561 m (Vadret da Morteratsch) from the AWS site altitudes. Air temperature was extrapolated to the AWS elevation with a constant temperature lapse rate Γ (Table 1). To obtain realistic accumulation

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at the beginning of the ablation season, we multiplied the CRU precipitation with a factor p (Table 3).

For the simulations on the global grid, the same temperature lapse rate Γ was used. Precipitation was extrapolated using two parameters; subsequent to multiplying the CRU precipitation by a factor p, we applied a linear increase in precipitation with altitude γ:

$$P_{\mathrm{ann}}(z) = pP_{\mathrm{ann,CRU}} + \gamma(z - z_{\mathrm{CRU}})\,. \tag{6}$$

4 Seasonal cycles at the AWS sites

The seven AWS sites cover a considerable range in climates, from an alpine climate with ablation dominated by solar radiation to a maritime climate with ablation all year round and an arctic climate with numerous snowfall events during summer. Since differences between the input data sets may affect the modelled mass balance, we shortly discuss the seasonal cycles of air temperature, diurnal temperature range and precipitation from the AWS and CRU data (Fig. 1). As the climatic conditions at Midtdalsbreen and S6 are similar to Storbreen and S5, respectively, these are not shown separately.

For Vadret da Morteratsch, Storbreen and Kongsvegen, the CRU temperatures correspond reasonably well with the AWS temperatures. Winter temperatures for Storbreen are lower in the CRU dataset, probably because weather stations are generally located in valleys where the air is less well mixed than on the glacier. Summer air temperatures for Kongsvegen are higher in the CRU data set, this could be due to local effects or the different measurement periods of the two datasets. For Breidamerkurjökull and S5 on Greenland, the seasonal temperature cycle is considerably larger in the CRU data set. Like on Storbreen, the air on the glacier is probably better mixed in winter, while it is katabatically cooled in summer. The diurnal temperature range in the CRU data set corresponds reasonably well with the AWS measurements

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on Vadret da Morteratsch and Kongsvegen, while it is considerably higher on the other three glaciers, especially in summer.

We performed model simulations for each of the AWS sites, using the three meteorological input data sets described in the previous section: AWS hour, AWS month and CRU. The calculations with AWS input data were carried out with values for τ, α_{ice}, c, ψ_{min} and T_{tip} calibrated for the particular sites (Table 2). With the CRU input data, three simulations were performed to examine the effect of the parameter values on the energy and mass balance. One simulation was done with the same parameter values as the runs with the AWS input data. The second and third runs were done with standard values, also used in the global simulations, using two different values for c (Table 2), representing the range of values found for the AWS sites. Measured and modelled seasonal cycles of net solar radiation, temperature-dependent flux and cumulative mass balance are shown in Fig. 2. Note that net solar radiation provides the majority of the melt energy at all locations and that the maximum contributions by net solar radiation and the temperature-dependent fluxes do not coincide at most of the AWS sites.

The simulations with hourly AWS data were carried out over all complete mass balance years (starting 1 October) with available data. Subsequently, the mean seasonal cycle was computed by averaging over all years for each calendar day, where the number of years varies for the different locations (Table **??**). With these realistic input data, the model is generally well able to capture the measured seasonal cycles in the energy fluxes.

Although we used measured incoming shortwave radiation for these simulations, net solar radiation depends on the albedo generated by the model. At all locations, modelled net solar radiation closely follows the measurements, demonstrating that the transition from snow to ice albedo is well represented in the model. Net solar radiation is too small in May for Vadret da Morteratsch, which is a direct result of the later disappearance of the snow cover in the model. On Kongsvegen, the variability in surface albedo and hence net solar radiation is difficult to model correctly. The AWS is situated approximately at the equilibrium-line altitude, where interannual variations in winter precipitation lead to net accumulation in some years and net ablation in others. Furthermore, frequent summer snowfalls cause large variability in the albedo. As we use the same amount of precipitation in each year, distributed equally over all days, the model cannot reproduce this variability.

Differences between the modelled and measured temperature-dependent fluxes are largest in spring. Although air temperatures are similar in spring and autumn, weather conditions are generally less humid, cloudy and windy in spring than in autumn. Both situations cannot be captured with a parameterization that depends on air temperature alone. Still, a significant part of the variations in measured ψ are reproduced with the model, indicating that using only temperature does not result in a large reduction of the variability.

Since we use a constant precipitation function, the build-up of the winter snowpack deviates from the observations at most locations. Despite these deviations, the onset of melt and the melt rate in summer correspond very well with the measurements. In general, we can conclude that the model parameterizations for surface albedo and ψ are applicable in this variety of climates.

When the input data is simplified to monthly AWS temperature data and calculated incoming solar radiation, the inter-daily variations disappear, but the overall shape of the seasonal cycles remains. At Breidamerkurjökull, S5 on Greenland and Kongsvegen, ablation events on warm days in winter and spring are not captured with the monthly data and the mass balance is less negative than observed. The summer melt rate is less sensitive to interdaily variations in temperature and is similar to the results obtained with hourly AWS input data at all sites.

Deviations from the measured seasonal cycles become larger when CRU data is used, especially for Breidamerkurjökull and S5 on Greenland. Since solar radiation is treated the same as with monthly AWS data, this is entirely the effect of the temperature input data. For these locations, the annual temperature cycle at the nearest CRU gridpoint is larger than measured at the AWS site (Fig. 1). In the summer months, temperatures are overestimated by 2 to 4 °C, which together with the large values for c

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lead to much larger temperature-dependent fluxes than measured. Too low winter temperatures at Breidamerkurjökull inhibit melting in winter, resulting in a best match with observed winter mass balance for $p = 0.0$ (Table 3). Although the seasonal cycle of the temperature-dependent fluxes also differs from the measurements at Vadret da Morteratsch and Storbreen, the annual mass balance does not deteriorate considerably when CRU input data is used.

The change from calibrated to standard τ is relatively small at all AWS sites and hardly affects net solar radiation and the surface mass balance. For Breidamerkurjökull and S5 on Greenland, the change to standard α_{ice} is large and has a considerable effect on net solar radiation and the surface mass balance. Replacing calibrated temperature-dependent flux parameters by set1 gives the smallest changes on Storbreen, where all parameters are rather close to the standard values. The change in the temperature-dependent flux on Vadret da Morteratsch and Kongsvegen is mainly attributable to the different value for T_{tip}, not c. For Breidamerkurjökull and S5, the substantially lower value for c compensates for the overestimated air temperatures in the CRU data, improving the match with measured temperature-dependent fluxes and ablation. With set2, the temperature-dependent flux barely changes for S5. For Breidamerkurjökull, the increase in ψ is balanced by a decrease in S_{net} due to the higher albedo and the mass balance for set2 is almost unchanged with regard to CRU data with calibrated parameters. The summer temperatures at the different sites clearly determine the sensitivity of the mass balance to changes in c; while the ablation increases by almost 6 m w.e. on Vadret da Morteratsch, the mass balance on Kongsvegen only changes by 0.5 m w.e. However, at both locations, this corresponds to a doubling of the surface lowering.

5 Altitudinal mass balance profiles

The precipitation multiplication factor p and the vertical gradient γ are expected to vary over the globe. Values for the two precipitation parameters were derived by a

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least-squares fit of the modelled to the observed winter mass balance profile for all glaciers with profiles available. Winter and annual mass balance profiles are available from the World Glacier Monitoring Service (WGMS, Haeberli et al., 2009, and earlier issues) for 80 glaciers, with a strong bias to Scandinavia and Canada (see Supplement for a table listing the 80 glaciers). For each glacier, the mean observed winter and annual balance profiles were calculated by averaging over all years with available mass balance measurements. Since the measurement date is seldomly reported for the winter balance, the mass balance calculated for 1 May (1 November on the Southern Hemisphere) is used in the fitting procedure.

Figure 3 shows the observed mass balance profiles for nine glaciers in different regions (Table 4), together with profiles simulated with calibrated (P-cal.) and standard (P-std.) precipitation parameters. The standard set of precipitation parameters was chosen in the midrange of the calibrated values: $p = 2.5$ and $\gamma = 1.5\,\mathrm{mm\,a^{-1}\,m^{-1}}$. For the other model parameters, we used the calibrated parameter set of the AWS site in the most similar climatic setting (Table 4). Similar to the previous section, we also show results obtained with a standard parameter set, to illustrate the effect of using different parameter values and to allow for a comparison between glaciers. A value of $c = 10.0\,\mathrm{W\,m^{-2}\,K^{-1}}$ (set1) was used, since this value generally gave good results for the AWS locations.

At most glaciers, the winter mass balance increases approximately linearly with altitude and can be fitted quite well by varying the two parameters. Repeating the fitting procedure with model parameters from set1 hardly affected the values found for p and γ, demonstrating the robustness of the fitting method and the small influence of melting on the winter balance profile.

The profiles modelled with standard parameters illustrate the necessity of calibrating the precipitation parameters. Reasonable agreement with the measured profiles is still obtained for Brewster Glacier, Storbreen and Austre Brøggerbreen, where the calibrated precipitation parameters are similar to the standard values. For the other glaciers, the modelled mass balance gradient and/or the absolute values of the winter

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balance are incorrect, resulting in large errors in the modelled annual mass balance. In the dry climates at Devon Ice Cap and Maliy Aktru, the overestimated precipitation gradient produces no solid precipitation at elevations below 850 m a.s.l. and 2600 m a.s.l., respectively, resulting in a sudden change in the annual mass balance gradient.

With calibrated values for p and γ, the measured annual mass balance gradient is captured by the model, although the absolute mass balance values often deviate from the observations (Fig. 3). This could be due to non-representative model parameters, but also to the extrapolation of the temperature data or the different periods represented by the climate data (1961–1990) and the measured profiles. For Storbreen, the only glacier with both an AWS site and a mass balance profile, the modelled mass balance is up to 1 m w.e. lower than the observations. This is likely due to the different periods represented by the CRU data (1961–1990), the AWS data (2002–2006) and the mass balance profiles available from WGMS (1990–2005). For most glaciers, locally measured meteorological data are not available and the cause for these discrepancies cannot be identified. We did not perform a multiple regression of a set of model parameters on the measured profiles, because this might result in an erroneous combination of parameter values with possible consequences for the mass balance sensitivity. Instead, we consider three cases:

1. The discrepancy is due to the modelled solar radiation, which can be calibrated by adjusting the value for τ.
2. The temperature-dependent flux is not correct and the parameters ψ_{min}, c and T_{tip} need to be modified.
3. Air temperatures T_a on the glacier are not correctly modelled, affecting ψ, the fresh snow albedo and the fraction of the precipitation falling as snow.

The real cause is likely a combination of these cases. By considering these three extreme cases separately, we can explore the effect a calibration could have on the mass balance sensitivities.

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For each of the 80 glaciers, the model was run with either varying τ (τ-cal.), the set of $\psi_{\min}$, c and T_{tip} (ψ-cal.) or T_{a} (T-cal.) and the best match was obtained from a least-squares fit to the mean measured annual mass balance profile. Since the altitudinal profiles modelled with set1 were very similar to the profiles modelled with parameter values from the most similar AWS climate, we only used the latter parameter set to perform the tuning to the measured annual balance. The optimized profiles are very similar for the three cases and generally in good correspondence with the measured altitudinal profile (Fig. 4). Since the adjusted parameters in some cases slightly affected the winter balance, we iteratively fitted the winter and annual mass balance until the best parameter combination was found. For the majority of glaciers, the final values for p and γ were very similar to the values already obtained.

The differences between the optimal precipitation parameter values for the three cases are generally small (Fig. 5), except for a few glaciers in Central Asia where large adjustments of the parameters were needed to match the observed mass balance profiles. For the majority of the glaciers, the amount of precipitation at the nearest CRU grid point is not sufficient to simulate the measured winter balance; p typically has a value of 2 or higher and a median value of 2.55. Values larger than 5 are rare and generally occur in combination with a zero or negative precipitation gradient. The median value of the calibrated precipitation gradient is 1.5 mm a^{-1} m^{-1}, with larger gradients on part of the Scandinavian glaciers and small or negative gradients on several glaciers in Central Asia and Central Europe. An example is Jamtalferner (Fig. 4), where the accumulation maximum occurs below the highest glacier elevation.

In Fig. 6, the optimal precipitation parameters for τ-cal. are shown versus the annual precipitation and the absolute latitude of the 80 glaciers. The annual precipitation is the area-averaged precipitation over the glacier, the absolute latitude represents the potential amount of solar radiation reaching the glacier surface. Although the calibrated values show some clustering per region, there is no apparent relationship with any of the climate variables.

The optimal values for τ all fall within the theoretically possible range $[0, 1]$ (Fig. 7) and can therefore solely explain the discrepancies in the modelled mass balance profiles. High values for τ are found for the glaciers in Central Europe and Northeastern Russia, indicating that the mass balance was initially overestimated. The opposite occurs for the glaciers in Central Asia, where the modelled mass balance was too negative. For the glaciers in Scandinavia and the Coast and Rocky Mountains, τ remains close to 0.5. A similar picture emerges if the annual mass balance is matched with a correction on the air temperature T_{corr}. When the parameters determining ψ are adjusted to match the observed mass balance profiles, the values are not as well constrained as for the other two cases. For some glaciers, unrealistically high values for c are found, always in combination with high values of T_{tip} and often low values of ψ_{min} (see Supplement). The model initially underestimated the measured mass balance at these glaciers, which is compensated by increasing T_{tip} and lowering ψ_{min}. As a result, air temperatures at these glaciers seldom exceed T_{tip} and c is not well constrained. Still, for roughly 80 % of the glaciers, c lies within the range found for the AWS sites.

The calibrated values of τ, T_{corr} and to some extent ψ_{min} increase with increasing annual precipitation and decrease with increasing continentality. This relation is contrary to the expected lower values for τ for humid climates with frequent cloud cover. Hence, it is more likely that ψ_{min} and/or T_{corr} need to be adjusted. For some of the glaciers in Central Asia, the elevation difference between the CRU grid point and the glacier altitude is large, increasing the potential error in the extrapolation of air temperature. Furthermore, no AWS data were available for this region to calibrate the temperature-dependent flux parameters; the lower values of ψ_{min} could be realistic.

6 Mass balance sensitivity

The calibrated mass balance profiles can be used to determine the sensitivity of the mass balance to changes in climatic variables. Of particular interest is the question whether the mass balance sensitivity depends on the variable calibrated to match the

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observed mass balance. In case τ is calibrated, the contribution of net solar radiation to the surface energy balance changes, while the temperature-dependent flux remains unchanged. On the other hand, when air temperature or ψ are adjusted, changes in the net solar radiation will be small. Supplementary to the commonly reported mass balance sensitivities to a 1 °C temperature change and a 10 % precipitation change, we compute the mass change induced by a 0.05 change in the atmospheric transmissivity τ. This is possible because net solar radiation is calculated separately from the other surface energy fluxes. Changes in τ of this magnitude for example correspond to observed interannual variability in incoming solar radiation caused by varying cloud conditions (e.g. Giesen et al., 2008), but also to decadal variations in irradiance related to global dimming and brightening (Ohmura, 2009; Wild, 2009).

We first discuss the changes in the altitudinal mass balance profiles, illustrated in Fig. 8 for the nine selected glaciers. The winter mass balance naturally changes at all glaciers when precipitation is increased or decreased. At maritime glaciers like Brewster and Koryto Glacier, the winter balance is also affected by changes in T and τ. The relative importance of changes in T, P and τ for the annual mass balance varies for the nine glaciers and with altitude. Generally, the largest changes occur for a 1 °C temperature change, while a 10 % change in precipitation has the smallest effect. The sensitivity to a 0.05 change in τ is often comparable to the sensitivity to a 10 % precipitation change, but is at some glaciers (Maliy Aktru, Peyto Glacier, Devon Ice Cap) as large as for a 1 °C temperature change. At all glaciers except Koryto Glacier, the mass balance in the ablation area is more sensitive to changes in the climatic variables, because of the change from snow to ice albedo that does not occur in the accumulation area.

The change in area-averaged annual mass balance induced by changes in temperature, precipitation and other climatic variables is often calculated to determine a glacier's sensitivity to climatic changes (e.g. Oerlemans and Fortuin, 1992; de Woul and Hock, 2005; Braithwaite and Raper, 2007). We calculated mass balance sensitivities for the 80 glaciers, combining the modelled profiles with each glacier's mean

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area-elevation distribution computed from the WGMS data. The mass balance sensitivities to changes in temperature C_T, precipitation C_P and atmospheric transmissivity C_τ are computed as (e.g. Oerlemans, 2001, p. 50):

$$C_T = \frac{B(T+1) - B(T-1)}{(T+1) - (T-1)} \tag{7}$$

$$C_P = \frac{B(P+10\,\%) - B(P-10\,\%)}{(P+10\,\%) - (P-10\,\%)} \tag{8}$$

$$C_\tau = \frac{B(\tau+0.05) - B(\tau-0.05)}{(\tau+0.05) - (\tau-0.05)}\,. \tag{9}$$

The mass balance sensitivities were calculated for the three calibration cases, the median values are listed in the Supplement. To put the variability in the mass balance sensitivities between the calibrated cases into perspective, we additionally calculated C_T, C_P and C_τ with uncalibrated model parameters (set1 with either standard or calibrated precipitation).

There is generally good correspondence between the sensitivities for the three calibrated cases (Fig. 9). The largest differences are found for C_T calculated with ψ-cal., for the glaciers with rather extreme values for the parameters determining ψ. For the case T-cal., unrealistically large values are obtained for C_P and C_τ on three glaciers in Central Asia. This is caused by a large negative anomaly in the mass balance, occurring when increased melt energy or reduced precipitation lets all solid precipitation disappear within a day, preventing the build-up of a snowpack. As already observed for the nine selected glaciers, C_τ has a value between C_T and C_P for most glaciers.

The variability in the mass balance sensitivities calculated with the same model parameters for all glaciers (set1, P-std.) is solely due to the different climates in which the glaciers are situated. Except for glaciers with precipitation parameters close to the standard values, the variability between the three calibrated cases is smaller than the effect of the precipitation calibration on the mass balance sensitivities (Fig. 9). The subsequent calibration to match the annual mass balance profile has the largest impact on C_T, the values for C_P and C_τ are not very different when set1 is used. For many glaciers in Central Asia, the mass balance sensitivity is highly dependent on the set of model parameters used.

As for the calibrated model parameters, we examine the dependence of the mass balance sensitivities on the annual precipitation and the absolute latitude. For most of the 80 glaciers, the glacier mass balance is not in equilibrium with the climate. We therefore imposed an additional temperature perturbation to obtain a zero area-averaged mass balance B_0 for all glaciers, using the model parameters for the case τ-cal. The mass balance sensitivities were recalculated using the B_0 configuration as the reference case. Except for the glaciers in Central Europe, where glaciers are far out of balance, the change in the mass balance sensitivities for the B_0 case is less than 0.1.

The sensitivity to temperature changes increases with increasing annual precipitation and is therefore highest for maritime glaciers in Scandinavia, New Zealand and Northeastern Russia. There is no apparent relation between C_T and latitude.

Similar to C_T, C_P is lowest for continental glaciers and high for maritime glaciers, with values generally being about half the value of C_T. The relation with annual precipitation is inherent to using a percentual change in precipitation: the higher the amount of precipitation a glacier receives, the larger the absolute change in precipitation. Like for C_T, there is no clear dependence of C_P on latitude.

The range of C_τ values is small compared to C_T and C_P, but shows a minor dependence on latitude, caused by the decrease in incoming solar radiation with increasing absolute latitude.

7 Conclusions and discussion

We applied a simple surface mass balance model to glaciers in different climates to explore its global applicability. The model uses a single expression for all temperature-dependent fluxes, combined with net solar radiation computed from incoming solar

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radiation and a surface albedo parameterisation. AWS records from glaciers in different regions were employed to calibrate the model parameters in the surface energy balance. Measured winter balance profiles for 80 glaciers were used to determine suitable values for the precipitation multiplication factor and vertical gradient. Further adjustments of the model parameters were necessary to match the observed annual mass balance profiles. For all 80 glaciers, we calculated the mass balance sensitivities to changes in temperature, precipitation and insolation and examined the effect of the model calibration on the obtained values.

The model used in this study contains a simplified computation of the surface energy fluxes, in order to be applicable when only air temperature and precipitation are available. By separating the contributions of solar radiation and all other fluxes to the energy balance, the effects of seasonal variations in both incoming solar radiation and temperature on the surface melt are included. This is especially important in regions where these seasonal cycles are asynchronous or where the annual amplitude in either insolation or air temperature is small. The relation between the temperature-dependent fluxes ψ and air temperature was derived from measurements. The model excludes variability in the surface energy balance caused by variations in wind speed, humidity and cloudiness. Still, the main characteristics of the seasonal cycle in ψ are captured with the model. This method is applicable to the majority of glaciers, where temperature changes govern the variations in the melt energy. We encourage other researchers to validate the proposed relation with their datasets. When it is not possible to measure or compute all the fluxes in the energy balance, the temperature-dependent flux can be estimated by subtracting the net solar radiation from the melt energy.

In the mass balance model, melting occurs whenever the surface energy balance is positive. In low-latitude regions where net solar radiation is large, this condition may also be met for air temperatures below the melting point. As long as the surface temperature is at the melting point, melt does happen in reality under these circumstances. For lower surface temperatures, ablation does not occur by melting but by sublimation of ice, which can be an important contributor to the total ablation (Mölg and Hardy, 2004; Wagnon et al., 1999). In our model, we do not distinguish between ablation by melting and sublimation, since the surface temperature is not calculated explicitly and information about humidity is not included. Sublimation is small on the majority of the glaciers, but should be taken into account when applying a mass balance model to low-latitude glaciers with a dry season.

The model performance relies for a large part on the climate data used as model input. Our results show that the effect of replacing hourly or daily meteorological data by monthly data is small when the accumulation and ablation seasons only slightly overlap. However, when significant melting takes place in winter or snowfall events frequently occur in summer, interdaily variations in temperature and preferably also precipitation should be included. Using climatic data that is not measured on the glacier itself increases the uncertainty in the modelled mass balance. Some typical features observed in the CRU data set used here are an overestimation of the annual and daily temperature range compared to the AWS data, but not for all locations. It is therefore impossible to identify a common cause or suggest a general correction method.

There is very large variation in the values found for the multiplication factor p and vertical gradient γ of precipitation, both within and between regions. At most glaciers, CRU precipitation needs to be increased by at least of factor of two to match the measured winter balance. The precipitation gradient takes both positive and negative values, with generally high values in maritime regions (Scandinavia, northeast Russia) and lower values for continental glaciers (e.g., Central Europe, Central Asia). While p depends on the climate data set chosen, γ should be more universally valid. The calibrated values for p and γ do not show a dependence on annual precipitation or latitude, which complicates the extrapolation of the precipitation parameters to glaciers without mass balance measurements.

Measured annual mass balance profiles could easily be matched by reasonable adjustments of either the atmospheric transmissivity, the parameters determining the temperature-dependent flux or the air temperature. Although the mass balance sensitivities to changes in temperature, precipitation and insolation are to a large extent determined by the climatic conditions, the calibration of model parameters can significantly affect the values. This implies that it is important to not only reproduce the measured mass balances, but to obtain representative model parameters as well. When no additional information is available, it is advisable to use values derived for a glacier in a similar climate and to make small adjustments to multiple parameters instead of choosing an extreme value for one of the parameters.

Our mass balance sensitivities to changes in temperature C_T are generally smaller than values obtained for the same glaciers in other global studies (e.g. Braithwaite et al., 2002; de Woul and Hock, 2005; Radić and Hock, 2011). These studies were all performed with a positive degree-day model, where the contribution of solar radiation to the surface energy balance is not considered separately. This difference in model setup may explain the lower values for C_T found in this study, but a possible other cause is a reduction of non-linear effects due to the low temporal (monthly) resolution of the climate data. For example, C_T found for Rembesdalsskåka in this study (–0.69 m w.e. K^{-1}) is much lower than the value found with a sophisticated mass balance model using hourly input data (–0.92 m w.e. K^{-1}, Giesen and Oerlemans, 2010). On the other hand, our values for C_P are often higher than in the studies referred to above, while for Rembesdalsskåka, it is comparable to the value found with the sophisticated model (0.32 and 0.31 m w.e. $(10\,\%)^{-1}$, respectively).

In addition to the mass balance sensitivities to temperature and precipitation changes, we calculated the sensitivity to perturbations of the atmospheric transmissivity that are comparable to observed interannual variability (due to different cloud conditions) and decadal variations (global dimming and brightening) in insolation. The resulting changes in the mass balance are in the same order of magnitude as caused by changes in precipitation and can be as large as the effect of temperature perturbations. This result illustrates the value of separately treating the contribution of net solar radiation to the surface energy balance from the other fluxes.

We can conclude that the mass balance model is applicable in regions with a climate similar to the locations it was calibrated for. By adjusting the model parameters within

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their uncertainty range, it is possible to match the measured annual mass balance profiles. However, differences with the measured mass balance may also be caused by non-representative climatic input data and therefore any change in the model parameters without further information is a mere guess. For glaciers with significant accumulation in summer, the model performance is not as good, but may be improved by using realistic temperature and precipitation data with daily resolution. The simulations point out that a correct representation of the winter precipitation is a prerequisite for obtaining realistic annual mass balance profiles and mass balance sensitivities.

Appendix A

A relation for ψ from measurements

To obtain a simple relation between T_a and ψ, we examined multi-annual records from automatic weather stations (AWSs). The Institute for Marine and Atmospheric research Utrecht (IMAU) operates AWSs on glaciers in a variety of climates, measuring all quantities needed for this analysis. We used data from IMAU AWSs in Switzerland, Norway, Iceland and Greenland; details of the AWSs and records are given in Table **??** and references therein. All records were analysed with the same energy balance model (e.g., see Van den Broeke et al., 2005; Giesen et al., 2008), solving the surface temperature from the surface energy balance with an iterative procedure. As the Arctic contains a large part of the total glacier area, we added a record from an AWS on Svalbard, analysed with the surface energy and mass balance model SOMARS (Krismer, 2009; Greuell and Konzelmann, 1994). In the tropics, the variability in the surface energy balance is determined by humidity instead of temperature changes. To verify whether a function for ψ only dependent on temperature is appropriate in these regions, AWS records from two tropical glaciers, Kersten Glacier in Tanzania and Zongo Glacier in Bolivia, were included in the analysis. Additionally, two short records from Arctic Canada and Southern Patagonia, covering only two months, were included. The

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surface energy fluxes on Kersten Glacier were calculated by Mölg et al. (2009), the other three datasets were analysed with the same energy balance model as used for the IMAU AWS records.

Since the temperature-dependent flux represents all energy fluxes other than net solar radiation, it was calculated as the sum of net longwave radiation and the turbulent fluxes of sensible and latent heat. As these fluxes also depend on cloudiness, humidity and wind speed, there is generally large scatter when plotting T_a versus ψ (Fig. 11a). Still, a pattern can be discerned, with ψ values close to zero for freezing temperatures and an approximately linear increase for T_a above the melting point. To obtain representative curves for each AWS site, we computed the mean value of ψ in each 1 °C temperature interval. Although the shapes of the curves vary due to the different local climates, some general features are found for all locations outside the tropics (Fig. 11b). For temperatures below the melting point, ψ is negative and varies only little with T_a. A continuous increase in ψ is seen for T_a above the melting point, with slopes varying for the different glaciers. The temperature-dependent flux increases more slowly with temperature for the smaller, more sheltered glaciers in the sample (Vadret da Morteratsch, Midtdalsbreen and Storbreen). The spread between the curves mainly results from general differences in wind speed, humidity and cloudiness between the sites. For example, average wind speeds on Midtdalsbreen are significantly higher than on Storbreen (Giesen et al., 2009) and Vadret da Morteratsch (Giesen et al., 2008), resulting in higher values for ψ. The curves for the two tropical glaciers have more negative values, since ψ needs to balance the much larger net solar radiative flux at low latitudes. Although for these glaciers the mean value of ψ also increases with T_a, such a relation does not emerge from the scatter plots and therefore has no solid physical basis.

Based on the relation generally found between ψ and T_a, we adopt a linearly increasing function with slope c for air temperatures above a threshold temperature T_{tip}. For temperatures below T_{tip}, we impose a constant value ψ_{min}:

$$\psi = \begin{cases} \psi_{min} + cT_a & \text{for} \quad T_a \geqslant T_{tip}\,; \\ \psi_{min} & \text{for} \quad T_a < T_{tip}\,. \end{cases} \tag{A1}$$

For every AWS location outside the tropics, we fitted a linear function to the increasing part of the ψ-curve, determined the minimum value ψ_{min} and calculated the corresponding value for T_{tip} (Table 2). The values obtained for Belcher Glacier and Glaciar Lengua only represent a short period in the ablation season and therefore only give a first estimate of typical values. Furthermore, ψ_{min} and T_{tip} could not be determined for Glaciar Lengua, since there is no tipping point in ψ for the available temperature range. The slope c varies largely between regions, but is similar for AWSs within a region (Storbreen and Midtdalsbreen in Southern Norway and S5 and S6 in Southwestern Greenland). The calibrated values for ψ_{min} and T_{tip} vary considerably between different sites as well and do not show any relation to other variables.

At low latitudes, humidity changes play an important role in the surface energy exchange (e.g. Kaser, 2001; Mölg et al., 2008). For example, on Zongo Glacier in Bolivia, our formulation of the temperature-dependent flux seems applicable in the humid season, while considerably lower values are attained in the dry season (Fig. 12). In such climates an energy flux relation dependent on both air temperature and relative humidity would be more appropriate, provided that humidity data are available.

Supplementary material related to this article is available online at: http://www.the-cryosphere-discuss.net/6/1445/2012/tcd-6-1445-2012-supplement.pdf.

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Table 1. Values for the fixed model parameters. The values for the parameters calibrated at the AWS sites are listed in Table 2.

Parameter	Symbol	Value	Unit
Latent heat of fusion	L_f	3.34×10^5	$\mathrm{J\,kg^{-1}}$
Water density	ρ_w	1000	$\mathrm{kg\,m^{-3}}$
Ice density	ρ_{ice}	900	$\mathrm{kg\,m^{-3}}$
Specific heat capacity of ice	c_{ice}	2090	$\mathrm{J\,kg^{-1}\,K^{-1}}$
Fresh snow albedo	α_{frsnow}	0.69–0.90	–
Firn albedo	α_{firn}	0.55	–
Albedo time-scale	t_*	21.9	days
Albedo depth-scale	d_*	0.001	m w.e.
Depth of subsurface layer	δz	2.0	m
Threshold temperature for snow	T_{snow}	1.5	°C
Temperature lapse rate	Γ	0.0065	$\mathrm{K\,m^{-1}}$

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Table 2. Values for the model parameters τ, α_{ice}, c ($W\,m^{-2}\,K^{-1}$), ψ_{min} ($W\,m^{-2}$) and T_{tip} (°C), calibrated for the individual AWS sites. The standard values used in the global application of the model are also listed. The values for the other model parameters are listed in Table 1.

Data set	Short name	τ	α_{ice}	c	ψ_{min}	T_{tip}
AWS site						
Vadret da Morteratsch	Mort	0.47	0.25	12.0	–26	+4.1
Glaciar Lengua	Leng	0.32	0.20	28.0	–	–
Midtdalsbreen	Midt	0.54	0.35	8.7	–25	–1.5
Storbreen	Stor	0.48	0.35	8.4	–19	+0.2
Breidamerkurjökull	Brei	0.44	0.15	22.6	–23	+0.9
S5 Greenland	GrS5	0.55	0.55	37.4	–26	+2.3
S6 Greenland	GrS6	0.63	0.55	40.5	–31	+1.1
Belcher Glacier	Belch	0.62	0.45	19.1	–42	+4.1
Kongsvegen	Kong	0.55	0.35	10.8	–33	–0.8
Standard						
Small c	set1	0.50	0.35	10.0	–25	+1.0
Large c	set2	0.50	0.35	30.0	–25	+1.0

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Table 3. Annual mean air temperature (T_{ann}), diurnal temperature range ($T_{range,AWS}$) and annual precipitation (P_{ann}) calculated from the AWS and CRU data. The annual precipitation at the AWS site ($P_{ann,AWS}$) is derived as described in the text. The CRU values are given for the gridpoint elevation (z_{CRU}), before applying the temperature lapse rate or the precipitation multiplication factor p.

AWS site	$T_{ann,AWS}$ (°C)	$T_{range,AWS}$ (°C)	$P_{ann,AWS}$ (m)	$T_{ann,CRU}$ (°C)	$T_{range,CRU}$ (°C)	$P_{ann,CRU}$ (m)	p –	z_{CRU} (m a.s.l.)
Vadret da Morteratsch	+1.6	6.5	2.4	–1.8	5.1	1.5	1.8	2676
Midtdalsbreen	–1.4	4.2	3.1	–2.6	5.2	1.2	2.1	1514
Storbreen	–1.9	4.3	2.6	–2.9	5.5	1.0	2.1	1466
Breidamerkurjökull	+2.0	4.1	1.7	+1.6	5.3	1.8	0.0	294
S5 Greenland	–5.5	4.9	0.5	–6.5	9.1	0.3	0.2	333
S6 Greenland	–10.7	6.6	0.4	–9.1	8.9	0.4	0.7	825
Kongsvegen	–9.0	4.2	1.0	–8.8	4.9	0.5	1.9	563

Table 4. Location (latitude, longitude), number of profiles (#), AWS parameter set, equilibrium-line altitude (ELA) and elevation of the nearest CRU grid point (z_{CRU}) for nine glaciers in different regions. Calibrated model parameters and mass balance sensitivities for all 80 glaciers can be found in the Supplement.

Glacier	Region	Latitude °N	Longitude °E	#	Set AWS	ELA (m a.s.l.)	z_{CRU} (m a.s.l.)
Djankuat Glacier	Caucasus	43.20	42.77	38	Mort	3213	2777
Brewster Glacier	New Zealand	–44.07	169.43	4	Stor	1923	1058
Jamtalferner	Central Europe	46.87	10.17	13	Mort	2965	2368
Maliy Aktru	Central Asia	50.08	87.75	12	Mort	3177	2747
Peyto Glacier	Coast/Rocky Mountains	51.67	–116.53	27	Mort	2720	2267
Koryto Glacier	Northeastern Russia	54.83	161.73	2	Stor	646	511
Storbreen	Scandinavia	61.57	8.13	16	Stor	1773	1466
Devon Ice Cap	Arctic	75.42	–83.25	3	Kong	1125	1489
Austre Brøggerbreen	Arctic	78.88	11.83	5	Kong	428	215

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Table 5. Location of the AWS sites and characteristics of the records used for model calibration and validation.

Location	Latitude (° N)	Longitude (° E)	Altitude (m a.s.l.)	Period	Interval (hour)	Reference
Kersten Glacier, Tanzania	–3.08	37.35	5873	09.02.05–23.01.08	1.0	Mölg et al. (2009)
Zongo Glacier, Bolivia	–16.25	–68.17	5060	28.01.05–27.01.06	0.5	Sicart et al. (2005)
Vadret da Morteratsch, Switzerland	46.42	9.93	2115	08.07.98–14.05.07	0.5	Oerlemans et al. (2009)
Glaciar Lengua, Gran Campo Nevado, Chile	–52.81	–73.00	450	23.02.00–12.04.00	1.0	Schneider et al. (2007)
Midtdalsbreen, Norway	60.57	7.47	1450	01.10.00–08.09.06	0.5	Giesen et al. (2008)
Storbreen, Norway	61.60	8.13	1570	06.09.01–11.09.06	0.5	Andreassen et al. (2008)
Breidamerkurjökull, Iceland	64.09	–16.32	190	06.05.02–06.05.06	0.5	unpublished
S5, K-transect, Greenland	67.10	–50.12	490	28.08.03–27.08.07	1.0	Van den Broeke et al. (2008)
S6, K-transect, Greenland	67.08	–49.38	1020	01.09.03–31.08.07	1.0	Van den Broeke et al. (2008)
Belcher Glacier, Devon Ice Cap, Canada	75.58	–81.43	500	02.06.08–31.07.08	1.0	unpublished
Kongsvegen, Svalbard	78.78	13.16	540	13.04.00–12.04.04	1.0	Krismer (2009)

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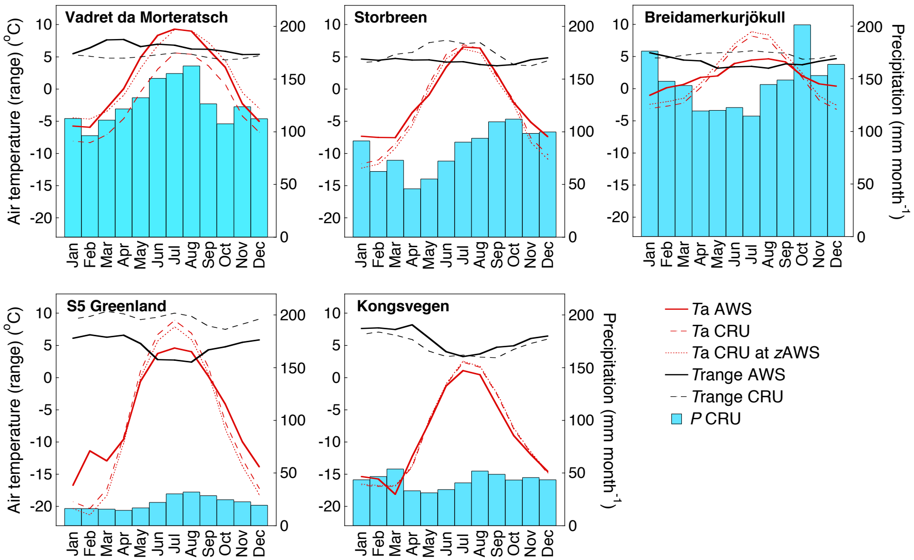

Fig. 1. Seasonal cycles of air temperature and diurnal temperature range from the AWS measurements and CRU data and precipitation from the original CRU dataset. To facilitate comparison, the temperature data from CRU have been extrapolated to the elevation of the AWS site using a lapse rate of $0.0065\,\mathrm{K\,m^{-1}}$.

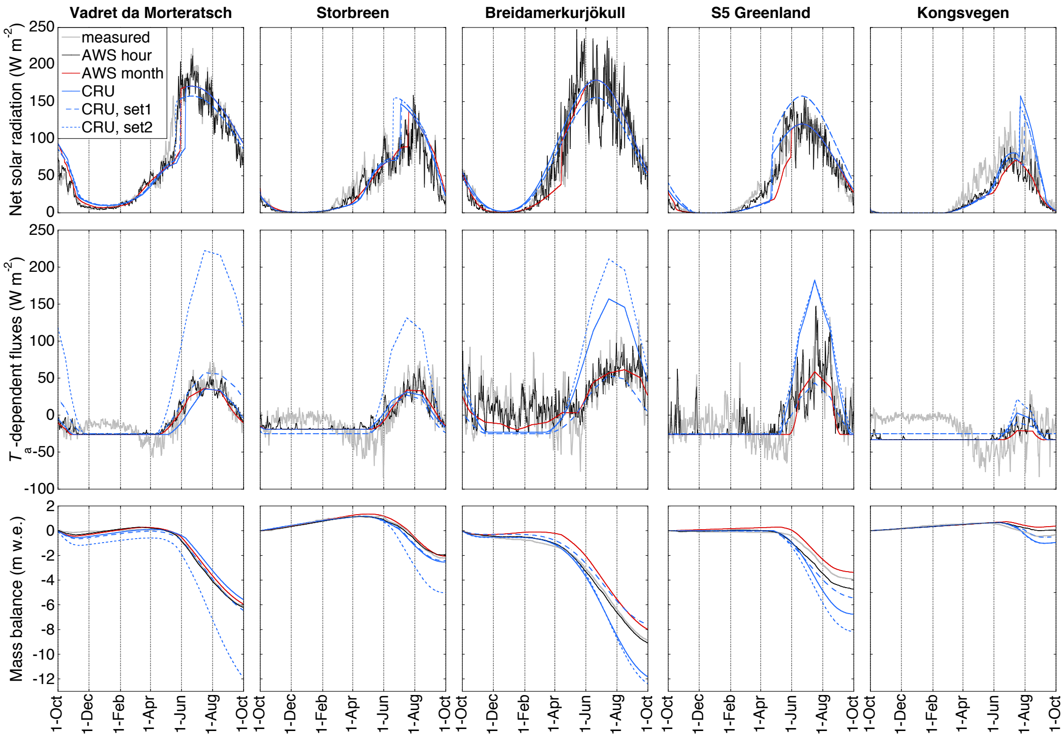

Fig. 2. Measured and modelled net solar radiation, temperature-dependent flux and mass balance at five AWS sites, using the three input data sets and calibrated model parameters. For the CRU input data, two simulations with standard values for τ, α_{ice} and c are also shown.

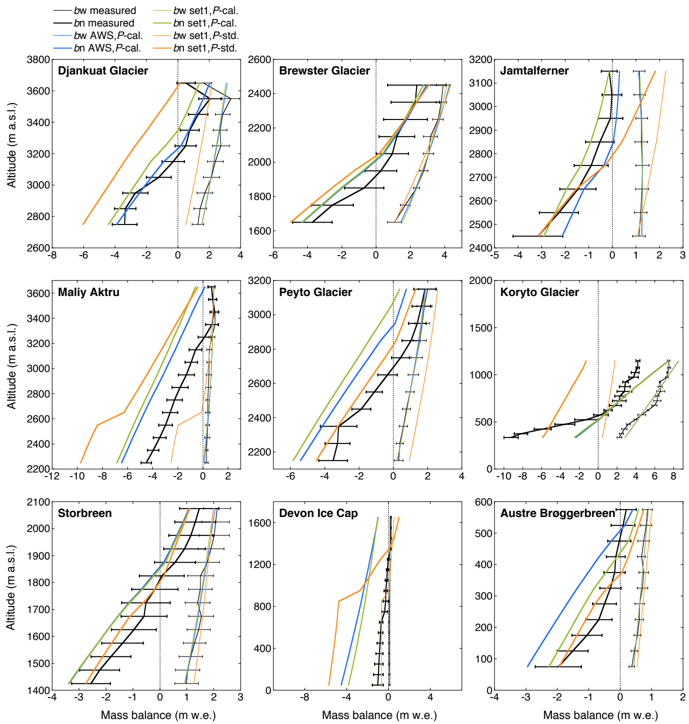

Fig. 3. Modelled and measured winter and annual mass balance profiles for glaciers in different regions (Table 4). The scales on the horizontal and vertical axes are chosen such that a 45° slope corresponds to a mass balance gradient of 1 m w.e. $(100\,\mathrm{m})^{-1}$ in all panels. The error bars on the measured profiles represent the standard deviation over the period of measurements. Model results with precipitation parameters calibrated with the winter balance profile (*P*-cal.) are shown with model parameters from the most similar AWS site (AWS) and standard values (set1). A simulation with standard precipitation parameters (*P*-std., $p = 2.5$ and $\gamma = 1.5\,\mathrm{mm\,a^{-1}\,m^{-1}}$) and set1 is also included.

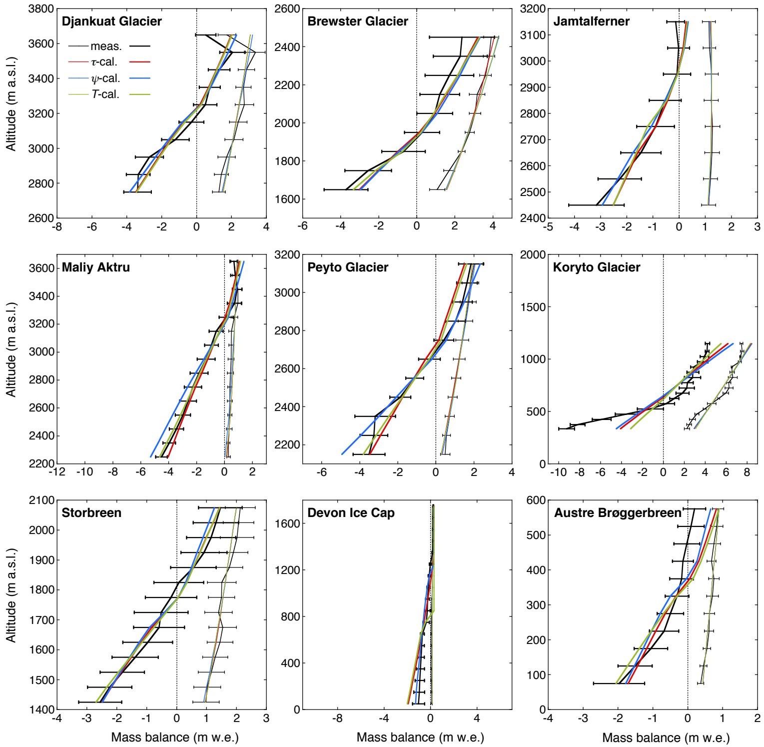

Fig. 4. Modelled and measured winter and annual mass balance profiles for glaciers in different regions (Table 4). The scales on the horizontal and vertical axes are chosen such that a 45° slope corresponds to a mass balance gradient of 1 m w.e. (100 m)$^{-1}$ in all panels. The error bars on the measured profiles represent the standard deviation over the period of measurements. Modelled profiles are shown for the three cases: τ-cal., ψ-cal. and T-cal.

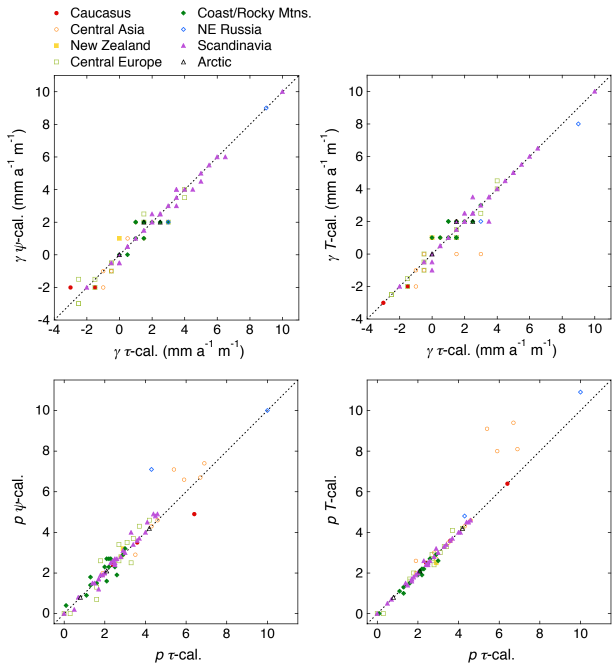

Fig. 5. Comparison of the vertical precipitation gradient γ and the precipitation multiplication factor p for the 80 glaciers for the three calibrated cases τ-cal., ψ-cal. and T-cal.

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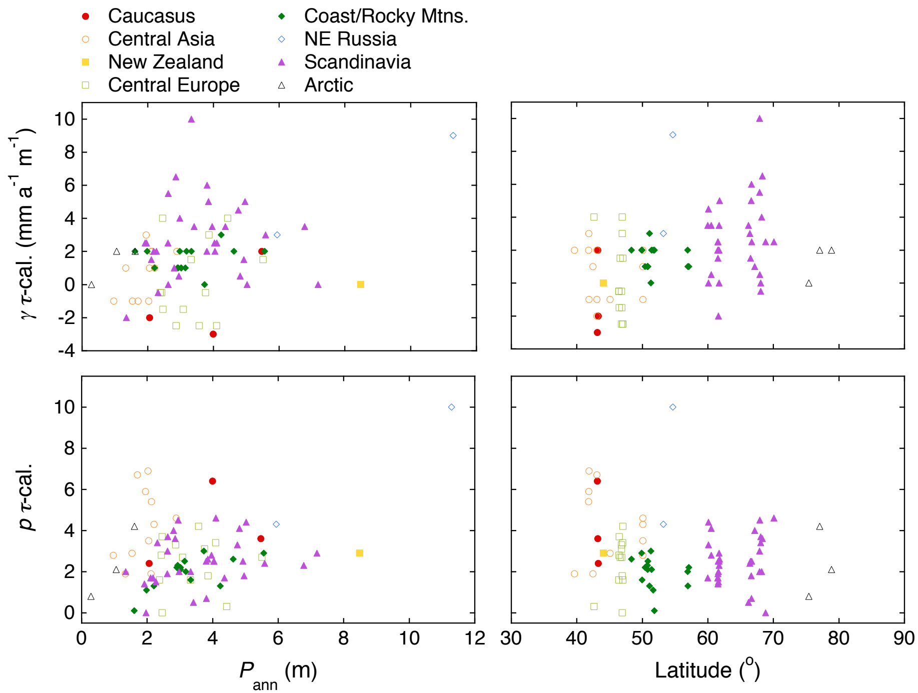

Fig. 6. Calibrated values for the precipitation parameters γ and p (case τ-cal.) versus annual precipitation P_{ann} and absolute latitude of the 80 glaciers.

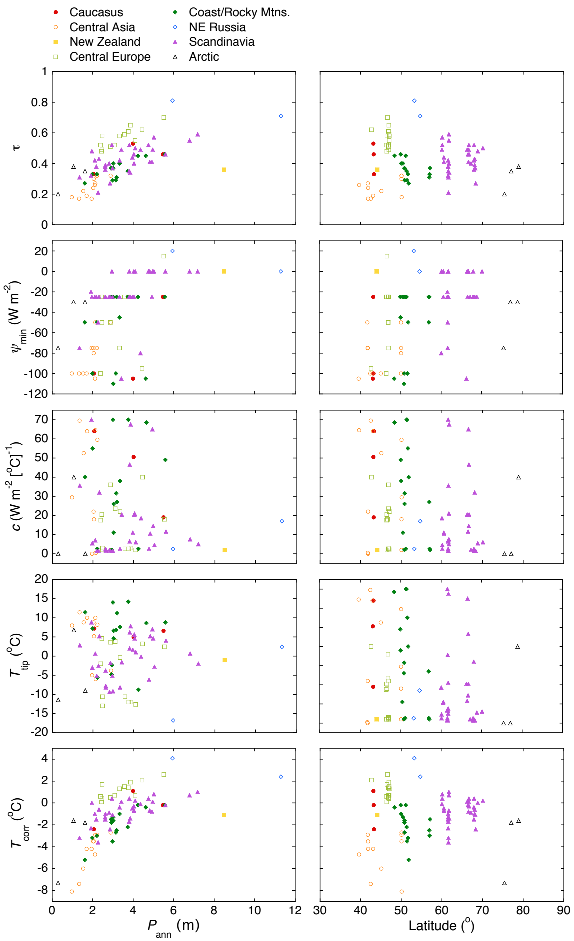

Fig. 7. Calibrated values for τ, $\psi_{\min}$, c, T_{tip} and T_{corr} versus annual precipitation P_{ann} and absolute latitude of the 80 glaciers.

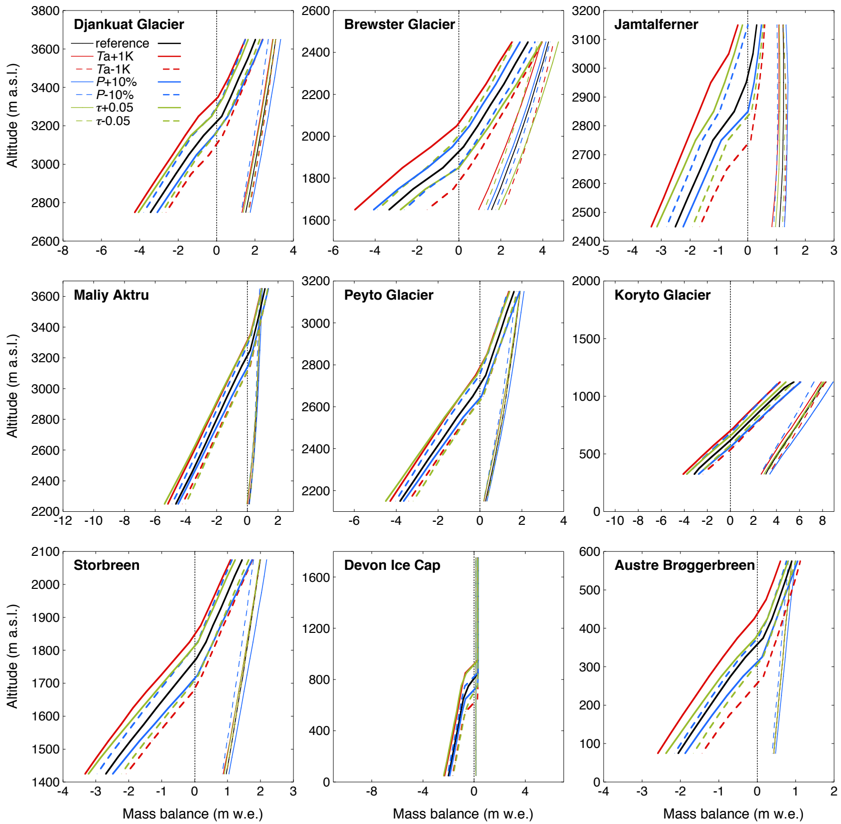

Fig. 8. Modelled winter and annual mass balance profiles for glaciers in different regions (Table 4), using the calibrated values for the annual air temperature (T-cal.) and additional perturbations of air temperature T, precipitation P and atmospheric transmissivity τ. The scales on the horizontal and vertical axes are chosen such that a 45° slope corresponds to a mass balance gradient of 1 m w.e. $(100\,\mathrm{m})^{-1}$ in all panels.

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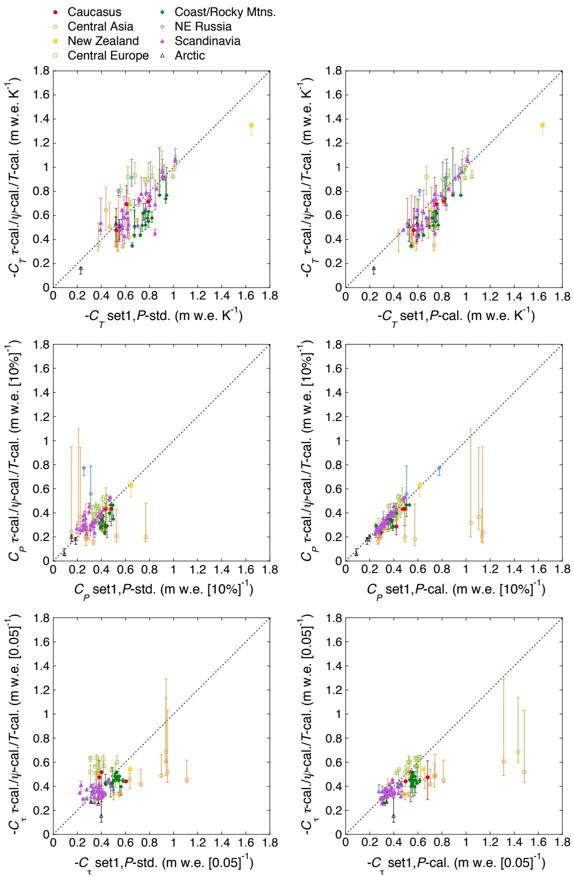

Fig. 9. Mass balance sensitivities to changes in air temperature C_T, precipitation C_P and atmospheric transmissivity C_τ calculated with standard parameters (set1) and standard (P-std.) or calibrated precipitation (P-cal.), compared to the values obtained for the three calibrated cases (τ-cal., ψ-cal. and T-cal.), shown as median, maximum and minimum values.

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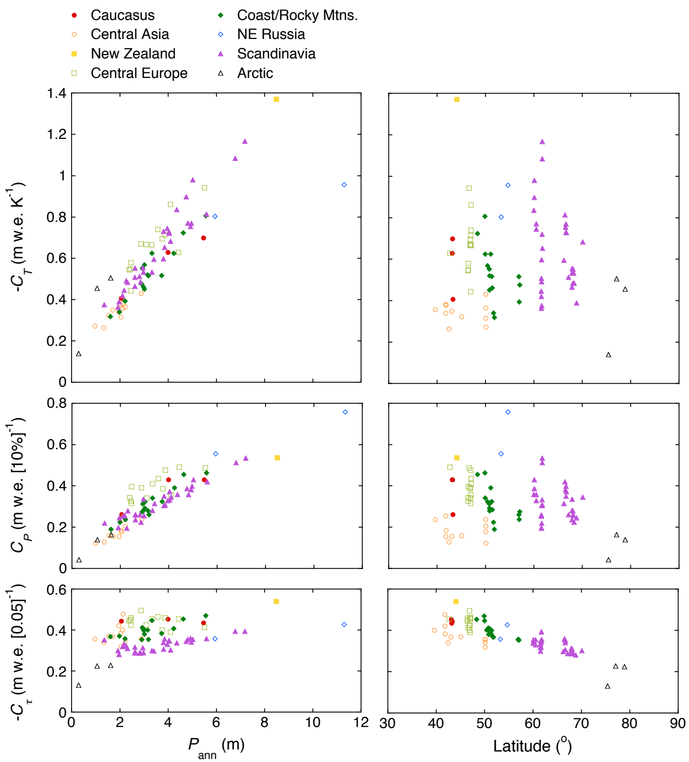

Fig. 10. Mass balance sensitivities to changes in air temperature C_T, precipitation C_P and atmospheric transmissivity C_τ for the case τ-cal. with area-averaged zero annual mass balance versus annual precipitation P_{ann} and absolute latitude for the 80 glaciers.

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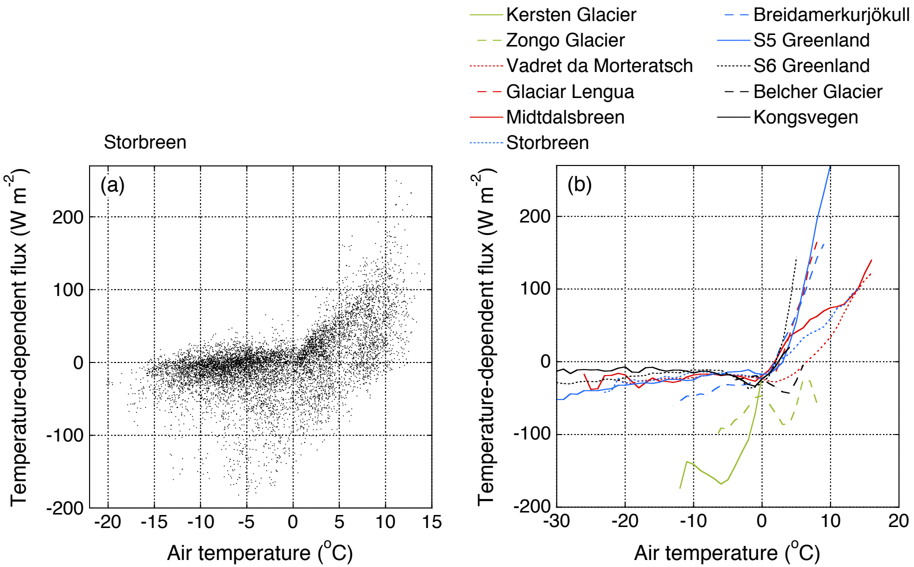

Fig. 11. The temperature-dependent flux ψ versus air temperature T_a with **(a)** hourly values for Storbreen for the year 2002 and **(b)** mean values for each 1 °C temperature interval for the eleven AWS records.

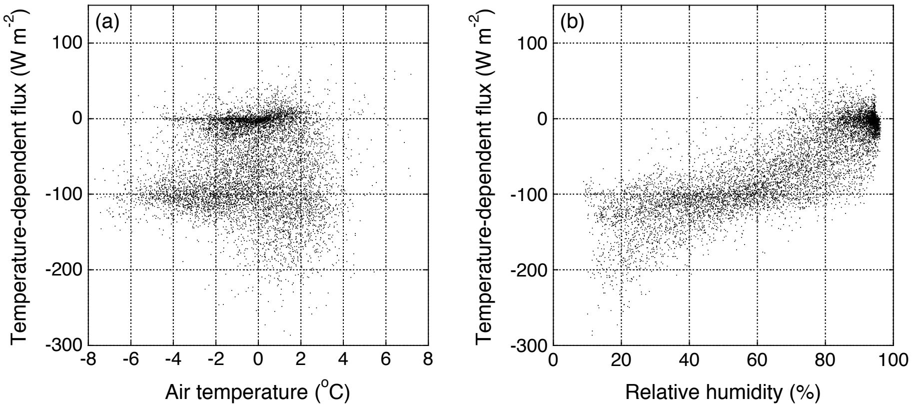

Fig. 12. The temperature-dependent flux versus **(a)** air temperature and **(b)** relative humidity for Zongo Glacier in Bolivia. Shown are hourly values for the period 28 January 2005 to 27 January 2006.

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