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2 m temperatures along melting mid-latitude glaciers, and implications for the sensitivity of the mass balance to variations in temperature

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ABSTRACT. In calculations of the variation in the 2 m temperature along glaciers, the lapse rate is generally assumed to be constant. This implies that the ratio of changes in the 2 m temperature above a glacier to changes in the temperature outside the thermal regime of that glacier ("climate sensitivity") is equal to 1. However, data collected during the ablation season on several mid-latitude glaciers show that this sensitivity is smaller than 1. The lowest measured value (0.3) was obtained on the tongue of the Pasterze, a glacier in Austria. The measured temperature distribution along the Pasterze cannot be described by a constant lapse rate either. However, there is almost a linear relationship between potential temperature and the distance along the glacier. This paper introduces a simple, analytical, thermodynamic glacier-wind model which can be applied to melting glaciers and which explains the observed "climate sensitivities" and temperature distributions much better than calculations based on a constant lapse rate.

This way of modelling the 2 m temperatures has implications for the sensitivity of the surface mass balance to atmospheric warming outside the thermal regime of the glacier. The magnitude of this sensitivity is computed with a surface energy-balance model applied to the Pasterze. When a constant lapse rate is used instead of the proposed glacier-wind model to compute changes in the 2 m temperature along the glacier, the negative change in mass balance due to 1°C warming is overestimated by 22%.

1. INTRODUCTION

In calculations of the variation in the temperature just above the surface along glaciers, the lapse rate is generally assumed to be constant (e.g. Oerlemans and Hoogendoorn, 1989; Jóhannesson and others, 1995; a constant lapse rate means that the change in temperature with elevation is constant). This assumption is made both for melting and for frozen surfaces. There are two reasons for reconsidering this assumption as far as melting glaciers are concerned.

First, measurements made during the melt season above the Pasterze, a glacier in Austria, show that the variation in the 2 m temperature along the glacier cannot be explained in terms of a constant lapse rate. In fact, the relation between potential temperature and the distance along the glacier is found to be almost linear (see Greuell and others, in press).

Secondly, the assumption of a constant lapse rate causes a problem in the calculation of changes in ablation due to changes in atmospheric temperature. Ablation changes are generally calculated by means of surface energy-balance models (e.g. Oerlemans and Hoogendoorn, 1989; Oerlemans and Fortuin, 1992) or degree-day models (e.g. Hoinkes and others, 1968; Jóhannesson and others, 1995). In most cases these models are forced by the temperature just above the glacier surface, typically at 2 m. Consequently, a change in temperature is also imposed at this level. If the lapse rate is assumed to be constant, the temperature change does not vary along the glacier and is equal to the temperature change outside the thermal regime of the glacier. Therefore, temporal variations in the 2 m temperature above the glacier are equal to those recorded at climate stations, and future changes in the 2 m temperature above the glacier are equal to those predicted by global atmospheric models. However, these equalities are not correct with respect to melting glaciers. Whereas the temperature of the free atmosphere (the part of the atmosphere not affected by the underlying surface) above a melting glacier varies, the temperature of the surface itself remains constant at 0°C. The 2 m temperature is intermediate between the temperature in the free atmosphere and the fixed temperature of the surface, and therefore the change in the 2 m temperature is smaller than that in the free atmosphere. Consequently, if a constant lapse rate is used to compute 2 m temperatures above the glacier from temperatures recorded at climate stations or predicted by atmospheric models, the sensitivity of ablation to variations in atmospheric temperature will be overestimated

The ideal solution to this problem is to use the temperature outside the thermal influence of the glacier as forcing and to compute melt by coupling a melt model to a mesoscale atmospheric model. The latter should extend beyond the thermal influence of the glacier and resolve details of the structure of the boundary layer above the glacier. However, such an approach is computationally expensive, and appropriate models still have to be developed. This paper provides an alternative approach in the form of a simple thermodynamic model of the glacier wind.

In section 2, we present the relevant data from the glacio-meteorological experiment on the Pasterze (PASTEX). In section 3, we introduce the glacier-wind model. This can be used to translate the temperature outside the thermal regime of the glacier to the 2 m temperature along a melting glacier, which can then serve as input to mass-balance models. In section 4 the PASTEX measurements are used to tune the glacier-wind model. Also, we describe some sensitivity experiments and investigate how well the model that was tuned with the temperature data from the Pasterze describes observations made on some other mid-latitude glaciers.

Finally, in section 5 the sensitivity of the surface mass balance of the Pasterze to atmospheric warming is computed by means of a surface energy-balance model. Experiments are described in which the change in the 2 m temperature along the glacier is computed both with the glacier-wind model and on the assumption of a constant lapse rate. The results of the two approaches will be compared.

2. MEASUREMENTS OF THE 2 M TEMPERATURE DURING PASTEX

In this section the relevant data from the glacio-meteorological experiment (PASTEX) on the Pasterze will be analyzed. During PASTEX, five energy-balance stations were established along the centre flowline of the Pasterze. Data were collected between mid-June and mid-August 1994. The surface profile of the glacier along its centre flowline is depicted in Figure 1, together with the location of the stations (U1–U5 and A1), directional constancies (q) and mean wind speeds (v). Two areas with relatively gentle slopes, namely, the upper part of the glacier and the tongue, are separated by a steep icefall. During PASTEX, the wind

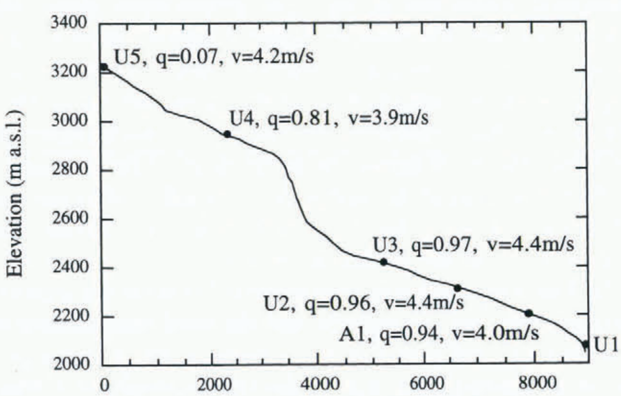

Fig. 1. Surface profile along the centre flowline of the Pasterze, with station names, directional constancies of half-hourly mean wind vectors (q) and mean wind speed (v) during PASTEX. Values for the individual stations cover exactly the same period, because directional constancy and mean wind speed were calculated from wind-direction and wind-speed samples that coincided with simultaneous measurements of the same variables in all the other stations. Without this restriction, the available datasets are longer, but this has little implication for the directional constancies and mean wind speeds. Only the directional constancy at U5 increases substantially, to 0.31. Data from U1 were not used in this paper since this station was located on the end moraine.

regime was dominated by the glacier wind, especially on the glacier tongue (Greuell and others, in press). This led to high values for the directional constancy of the wind vector (0.94–0.96 on the tongue). The mean wind speed hardly varied along the glacier.

In the present paper we have used the measurements of the 2 m temperature. Two datasets were extracted: the variation in the mean temperature along the glacier, and the variation in "climate sensitivity" along the glacier. "Climate sensitivity" is defined here as the ratio of changes in the 2 m temperature above the glacier to changes in the temperature outside the thermal regime of the glacier. In this paper, temperatures measured on the summit of Sonnblick (3106 m a.s.l.), a mountain located some 20 km east of the Pasterze, are assumed to represent conditions outside the thermal regime of the glacier. The validity of this assumption will be discussed in section 4. Figure 2a depicts the mean observed temperature during PASTEX as a function of elevation, together with the mean temperature on Sonnblick during the same period. The data from Sonnblick, U4 and U5 show that temperatures above the glacier are relatively low. This is due to the cooling effect of the glacier. More remarkably, the usually assumed constant lapse rate is not visible. The temperature at U3 is even higher than the temperature 215 m lower at A1, and the temperature at U5 (3225 m a.s.l.) is almost equal to that at U4 (2945 m a.s.l.). A much better description of the observed temperature distribution is given by a linear relation between potential temperature and the distance along the glacier (Fig. 2b).

Next, the "climate sensitivity" during PASTEX is considered. First, it was determined for each of the five mast locations on the glacier. Figure 2c gives an example (A1, 2205 m a.s.l.). Daily mean temperatures at A1 are plotted against the daily mean temperature on the summit of Sonnblick (3106 m a.s.l.), a mountain located approximately 20 km east of the Pasterze. A line was fitted using the least-squares method, its slope yielding the local "climate sensitivity". Then, the values of the local "climate sensitivity" were plotted against the distance along the glacier (Fig. 2d). As expected, the "climate sensitivity" is smaller than 1 at all locations and decreases with distance along the glacier.

3. MODEL OF THE TEMPERATURE DISTRIBUTION

In this section we discuss a model describing the variation in the 2 m temperature along the flowline of a melting glacier. A schematic diagram of the model is shown in Figure 3. The first assumption is that there is a glacier-wind layer above the surface. The glacier wind is a drainage wind that is due to the cooling of air with a temperature higher than 0°C over a melting glacier. Experiments (see e.g. Obleitner, 1994; Greuell and others, in press) show that the glacier wind often dominates the wind regime over melting glaciers. In the model, parcels travel down along the glacier within the glacier-wind layer. It is assumed that the parcels travel parallel to the surface with a speed u and have a temperature T. Vertical variations in these variables within the glacier-wind layer are not considered. The thickness of the layer normal to the surface is denoted by H. In the coordinate system, the elevation is given by z, and the distance along the horizontal projection of the flowline by x, with $x = 0$ at the top of the flowline. From now on, this distance

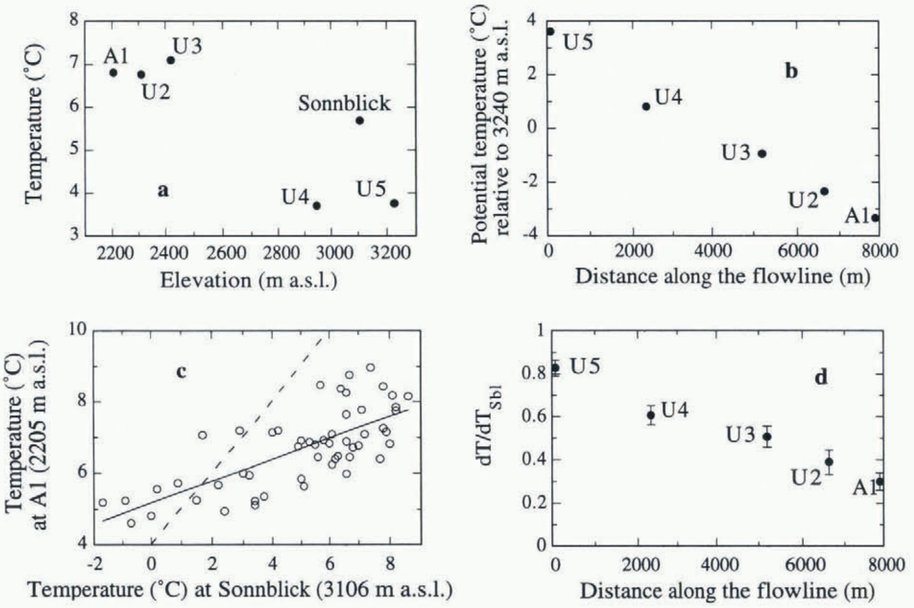

Fig. 2. Data concerning the 2 m temperature during PASTEX as used for this study. The mean temperature distribution during the experiment is plotted as temperature against elevation (a) and potential temperature against the distance along the flowline (b). (c) is a scatter plot of the daily mean temperature at A1 against the daily mean temperature at Sonnblick. The dashed line is an arbitrary 1:1 line, and the solid line shows the best linear fit using the least-squares method. Its derivative ("climate sensitivity") with error margin for the five glacier stations is plotted against the distance along the flowline in (d).

will simply be referred to as the distance along the flowline. The glacier has a slope angle α.

For the analysis, one needs a relation between potential temperature (Θ) and the temperature T. If the slope is constant, and potential temperature is taken with respect to the pressure prevailing at $x = 0$, the relation reads:

$$T = \Theta + bx \tag{1}$$

$$\text{with} \qquad b = \Gamma_d \tan \alpha \tag{2}$$

where Γ_d is the dry adiabatic lapse rate (-0.0098 K m^{-1}).

The temperature of the model parcels changes due to two processes, namely, adiabatic heating and exchange of sensible heat with the underlying surface. Therefore, temperature changes due to entrainment, phase changes, radiation divergence and variation of fluxes in the horizontal direction normal to the flowline are neglected. This two-term balance in the heat budget was found by Van den Broeke (in press) for site A1 during PASTEX. Since adiabatic heating does not affect potential temperature, the rate at which the potential temperature of the parcels changes is determined only by the sensible heat flux:

$$H \frac{d\Theta}{dt} = -c_H (T - T_s) u \tag{3}$$

where c_H is the bulk transfer coefficient for heat, which depends on the surface roughness lengths and the stratification (see Stull, 1988). Further, T_s is the surface temperature (0°C). Substituting $dx = u \cos\alpha \, dt$ and rearranging Equation (3) yields:

$$L_R d\Theta = -T dx \tag{4}$$

$$\text{where} \qquad L_R = \frac{H \cos \alpha}{c_H} \tag{5}$$

and T is given in °C. Note that the wind speed has disappeared from the equation. Next, it is assumed that H/c_H is constant along the flowline, so for a constant slope L_R is a constant, too. After combining Equations (1) and (4), we can solve the resulting differential equation with the assumption of a constant L_R:

$$\Theta(x) = (T_0 - T_{eq}) \exp\left(-\frac{x}{L_R}\right) - bx + T_{eq} \tag{6}$$

or in terms of temperature:

$$T(x) = (T_0 - T_{eq}) \exp\left(-\frac{x}{L_R}\right) + T_{eq} \tag{7}$$

where $T_{eq} = bL_R$ and the temperature T_0 at $x = 0$ is inserted as a boundary condition.

The assumption of a constant H/c_H is debatable. Ohata (1989) analyzed the glacier wind with a model in which entrainment was incorporated. His numerical solution suggests that H is almost proportional to the square root of x, which is in agreement with what is generally known about the growth of internal boundary layers (Stull, 1988). On the other hand, by analyzing the PASTEX data, Van den Broeke (in press) found that entrainment is weak over the tongue of the Pasterze, implying that variations in H with x should be small. The model equations discussed here can also be

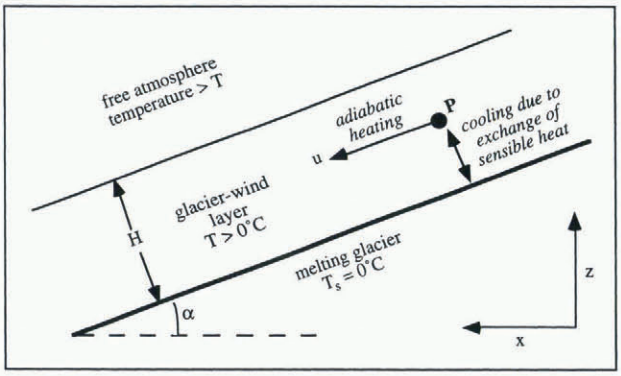

Fig. 3. Schematic diagram of the model. The large dot P represents the parcel considered in the model.

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solved analytically by taking H/c_H proportional to the square root of x. However, the observations were found to be less well described by taking H/c_H proportional to the square root of x than by taking H/c_H as a constant.

The temperature distribution of Equation (7) is plotted in Figure 4. The temperature approaches an equilibrium value (T_{eq}) in an exponential way with length scale L_R. Equilibrium is reached when adiabatic heating balances cooling due to exchange of sensible heat with the glacier. Note that because the equilibrium temperature is proportional to $\sin\alpha$, temperatures over steeper glaciers tend to be higher. For $x \gg L_R$ the temperature does not depend on the temperature of the parcels when injected into the glacier-wind layer at $x = 0$. This means that the temperature for $x \gg L_R$ is independent of the temperature outside the thermal influence of the glacier. These regions therefore are not affected by a climatic change. In fact, the influence of variations in T_0 decreases exponentially along the glacier according to:

$$\frac{\mathrm{d}T(x)}{\mathrm{d}T_0} = \exp\left(-\frac{x}{L_R}\right). \qquad (8)$$

In the case of PASTEX, the choice of the location of the top of the flowline was obvious since there was a well-defined crest. However, in other cases the choice may be rather arbitrary. Also, even if it is easy to define this point, the glacier may have a thermal influence at $x = 0$. Figure 2d illustrates that this occurred during PASTEX. Temperature fluctuations at U5 (at $x = 0$) were smaller than simultaneous temperature variations outside the thermal influence of the glacier ($\mathrm{d}T/\mathrm{d}T_{Sbl} = 0.83$; see Fig. 2d). This effect may be connected with the fact that the assumption that a glacier wind is always present along the entire glacier is incorrect. Although, during PASTEX, glacier winds dominated along the glacier from A1 to U4, resulting in high values of the directional constancy (0.81–0.97), the wind was much more variable at U5, where the directional constancy was only 0.07 (see Fig. 1). Thus, at this location the wind direction was often upwards along the flowline, which explains the influence of the melting glacier on the measured 2 m temperatures there. Moreover, air parcels moving across the crest onto the Pasterze may have been affected by snowfields

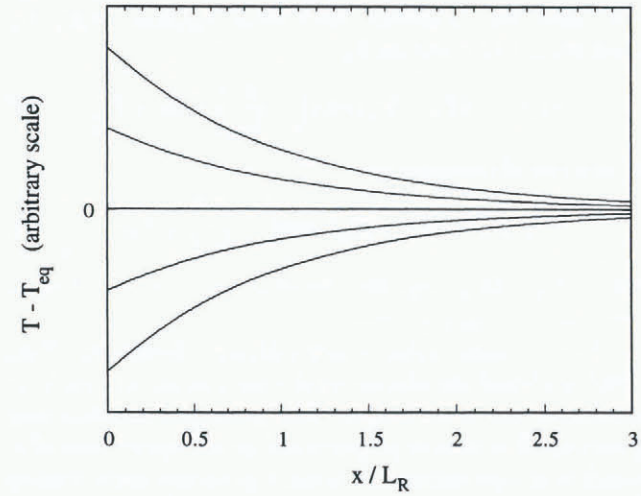

Fig. 4. Variation in the temperature of the glacier-wind layer (T) along the flowline of a melting glacier ($x = 0$ at the top of the flowline) according to the model treated in this section. The five curves are for different values of $T_0 - T_{eq}$.

on the weather side. In order to take a non-zero "climate sensitivity" at the top of the flowline into account, an unknown distance x_0 is added to the coordinate x, so Equations (6) and (7) become:

$$\Theta(x) = (T_0 - T_{eq})\exp\left(-\frac{x + x_0}{L_R}\right) - b(x + x_0) + T_{eq} \qquad (9)$$

$$T(x) = (T_0 - T_{eq})\exp\left(-\frac{x + x_0}{L_R}\right) + T_{eq}\,. \qquad (10)$$

Here T_0 is the temperature at $x = -x_0$, which is the imaginary location where the parcels enter the glacier-wind layer and potential temperature is taken relative to the elevation z_0 at $x = -x_0$. Since $x = -x_0$ is an imaginary location, one has some freedom in the choice of z_0. Constraints on this choice will be discussed in the following section. The temperature T_0 can be estimated from the temperature measured at a climate station (T_{cs}):

$$T_0 = T_{cs} - \gamma\,(z_{cs} - z_0) \qquad (11)$$

where γ is the lapse rate outside the thermal influence of the glacier and z_{cs} is the elevation of the climate station. Note that the climate station has to be located outside the thermal influence of the glacier. From Equations (10) and (11) the "climate sensitivity" is derived as:

$$\frac{\mathrm{d}T(x)}{\mathrm{d}T_{cs}}\exp\left(-\frac{x + x_0}{L_R}\right) \qquad (12)$$

Temperature and "climate sensitivity" along the flowline are described by Equations (10)–(12). These equations contain five unknown parameters, x_0, L_R, b, z_0 and γ.

4. TUNING AND VALIDATION OF THE MODEL

In this section, the values of the unknown parameters will be determined from the PASTEX data. Also, we investigate how representative the conditions at Sonnblick are for conditions outside the thermal influence of the glacier. Finally, the model tuned with the PASTEX data will be validated by means of several datasets from other mid-latitude glaciers.

Therefore, first the values of the unknown parameters were determined for the Pasterze. In view of the interest in the climate sensitivity of glaciers, priority was given to obtaining the correct "climate sensitivity", rather than the correct temperature distribution. Therefore, a fit of the exponential function Equation (12) was made to the "climate sensitivities" determined for the individual mast locations (Fig. 5; cf. Fig 2d), yielding values of x_0 (1440 ± 630 m) and L_R (8340 ± 960 m). Since $H = c_H L_R \cos^{-1}\alpha$ (see Equation (5)), the thickness of the glacier-wind layer can be estimated. With $c_H = 0.002$ (from Stull, 1988, table 7-3; valid for 10 m winds over snow surfaces) and $\tan\alpha$ equal to the mean slope between U5 and A1 (= 0.13), $H = 17$ m. This is close to the thickness (20 m) of the cooled layer in the mean profile at A1 (2205 m a.s.l.), as obtained from upper-air soundings (Van den Broeke, in press).

Next, the observed mean 2 m temperature distribution along the glacier (Fig. 2a and b) was considered. In order to simulate this distribution with the model, the mean temperature on Sonnblick ($z_{cs} = 3106$ m a.s.l.) during PASTEX ($T_{cs} = 5.7°$C) was inserted into Equation (11). The parameter γ was set equal to the mean lapse rate above the glacier-wind layer according to the upper-air soundings at A1 (-0.007 K m^{-1}), and the elevation z_0 at $x = -x_0$ was taken equal to the elevation at the top of the flowline

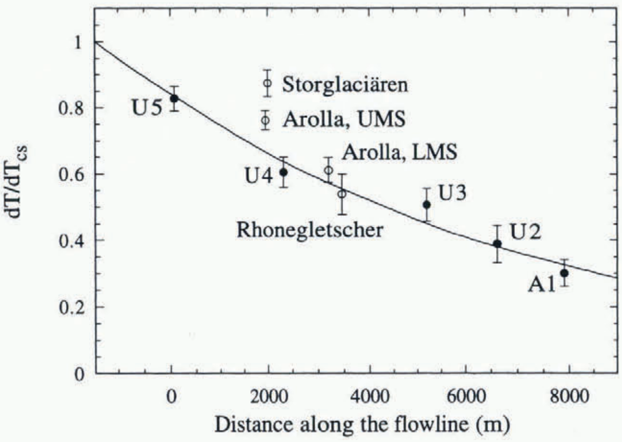

Fig. 5. Variation in the sensitivity of the local temperature to the temperature outside the thermal influence of the glacier (see Fig. 2d) along the glacier. Values were derived from data collected on Storglaciären, Rhonegletscher (one station each), Haut Glacier d'Arolla (two stations) and the Pasterze (A1, U2, U3, U4 and U5). The best fit of Equation (12) to the Pasterze data using the criterion of least squares is given by the continuous line. Error bars give standard deviations of the estimates of the slopes of the regression lines at the individual stations (see Fig. 2c).

(3240 m a.s.l.). Consequently, $T_0 = 4.8°C$. The next step was to compute the potential temperature as a function of x by inserting the values of T_0 and b into Equation (9). The parameter b was obtained by means of Equation (2) with:

$$\tan\alpha = \frac{z_0 - z(\mathrm{U1})}{x(\mathrm{U1}) + \mathrm{x}_0} \tag{12}$$

so $b = 1.10 \times 10^{-3}$. Therefore, $\tan\alpha$ is the mean slope between $x = -x_0$ and the terminus of the glacier. The result is shown by the curve "constant b (Sonnblick)" in Figure 6. Next, $\Theta(x)$ was converted into temperature by means of:

$$T(x, z) = \Theta(x) - \Gamma_\mathrm{d}[z(x = 0) - z(x)] \tag{13}$$

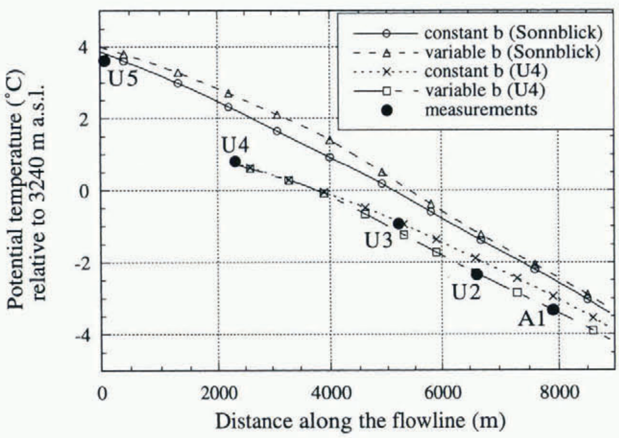

Fig. 6. Calculated and measured 2 m potential temperature along the glacier during PASTEX. The measurements are mean values for approximately 48 days. The calculations use the mean temperature on Sonnblick or at U4 as a boundary condition. The equations are solved numerically so variations in slope along the glacier can be taken into account ("variable b"), or they are solved analytically ("constant b"). In the latter case the mean slope between $x = -x_0$ and the terminus is used to compute the parameter b.

where $z(x)$ is the surface profile of the glacier (Fig. 1). Note that by means of this procedure the solution in terms of $T(z)$ becomes different from the one that would have been obtained directly from Equation (10), because Equation (10) neglects variations in slope along the glacier. Temperature is shown as a function of elevation by the curve "constant b (Sonnblick)" in Figure 7.

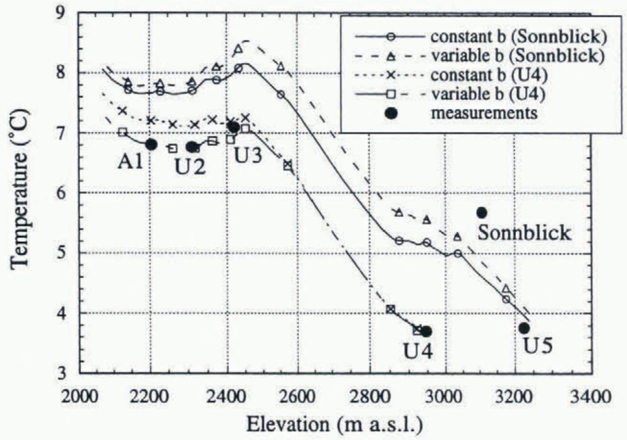

Fig. 7. Calculated and measured 2 m temperature vs elevation during PASTEX. The measurements are mean values for approximately 48 days. The calculations use the mean temperature on Sonnblick or at U4 as a boundary condition. The equations are solved numerically, so variations in slope along the glacier can be taken into account ("variable b"), or they are solved analytically ("constant b"). In the latter case the mean slope between $x = -x_0$ and the terminus is used to compute the parameter b.

Before simulated and observed temperature distributions are compared, the effect of a constant slope is discussed. Both the parameters b (Equation (2)) and L_R (Equation (5)) are a function of the slope. However, since for small slope angles local variations in slope have a much larger effect on b (proportional to $\tan\alpha$) than on L_R (proportional to $\cos\alpha$), only the effect of a constant b is considered here. This is done by comparing the solution for a constant b with a solution that takes into account variations of b along the glacier. The latter was found by numerical integration of Equations (1) and (4) along the flowline so that the variation of b with x could be taken into account. This solution can be considered the best approximation of the exact solution of the model equations for a glacier with variable slope. It is shown in Figures 6 and 7 by the curves "variable b (Sonnblick)". Since there is little difference between the "constant b" and "variable b" solution, it is concluded that the assumption of a constant b is allowed and that the equations can be solved analytically, at least for the Pasterze.

Both solutions describe the shape of the observed temperature distribution rather well, except for the upper part of the glacier between U4 and U5 (Figs 6 and 7). However, temperatures are systematically too high at all stations except U5, but, as shown by the curves "constant b (U4)" and "variable b (U4)", better agreement can be achieved if U4 is taken as the point where the parcels enter the glacier-wind layer. This situation was simulated by setting $T_0 = T(\mathrm{U4})$, $x_0 = -x(\mathrm{U4})$, $z_0 = z(\mathrm{U4})$ and by calculating b from the mean slope between U4 and the terminus of the

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glacier. It is concluded that the model gives a fair description of the thermodynamics of the glacier-wind layer downwards from U4, but it fails upwards from this location. Indeed, the glacier wind was less frequent in the upper part of the glacier, as demonstrated by the directional constancies in Figure 1. Also, one can speculate that on occasions if there is a glacier-wind layer in the upper part of the glacier the layer thickness will be smaller than further down the glacier owing to the small fetch (see Ohata, 1989). Such a relatively shallow layer might qualitatively explain the lack of warming between U5 and U4, because the cooling rate due to the exchange of sensible heat with the surface increases with decreasing thickness of the layer (Equation (3)).

Figure 8 shows results of some experiments that we performed in order to study the effect of variations in b and z_0 (variation of γ within its range of uncertainty had little effect on the simulated temperature distribution). The curve "constant slope (Sonnblick)" from Figure 7 was used as a reference. It should be recalled here that this curve was computed from the temperature on Sonnblick, from a constant b derived from the mean slope between $x = -x_0$ and the glacier terminus and from a z_0 equal to the elevation at $x = 0$ (3240 m a.s.l.). The corresponding temperature distribution calculated with variable b is also shown in the figure. In the first sensitivity experiment, the slope used to calculate b was taken to be equal to the mean slope between $x = 0$ and the terminus instead of the mean slope between $x = -x_0$ and the terminus. The reference solution was found to approach the "variable b" solution more closely than the solution found when b was computed as in this sensitivity experiment. Therefore, if one bears in mind that the "variable b" solution is the optimal solution of the model equations, the better option, at least in this case, is to compute b from the mean slope between $x = -x_0$ and the terminus. In two other experiments the effect of variations in z_0 was studied. If z_0 is considered as an elevation within a distance x_0 from the top of the flowline but on the glacier, it might range roughly between 3100 and 3400 m a.s.l. Two temperature distributions were computed with these extreme values of z_0 and a constant b with the same value as in the reference experiment. Temperatures were found to be just over 0.5°C lower ($z_0 = 3100$ m a.s.l.) and higher ($z_0 = 3400$ m a.s.l.) along the entire flowline. It is concluded that z_0 might be used for tuning the absolute temperatures. However, even if $z_0 = 3100$ m a.s.l., the shift in the calculated temperature is insufficient to explain the temperatures measured downwards from U4. Moreover, as shown before, the discrepancy between measurements and calculations in the reference run may well be attributed to the fact that the glacier-wind assumption does not hold above U4. Therefore, tuning the model with z_0 might lead to a better overall agreement between measurements and calculations for the wrong reasons, while at the same time the simulation of the pattern below U4 gets poorer. Consequently, a correction that is constant with elevation will be used to tune the glacier-wind model (see next section). Note that the climate sensitivity $\mathrm{d}T/\mathrm{d}T_{cs}$ (Equation (12)) is not affected by either of these tuning procedures, because it does not depend on z_0.

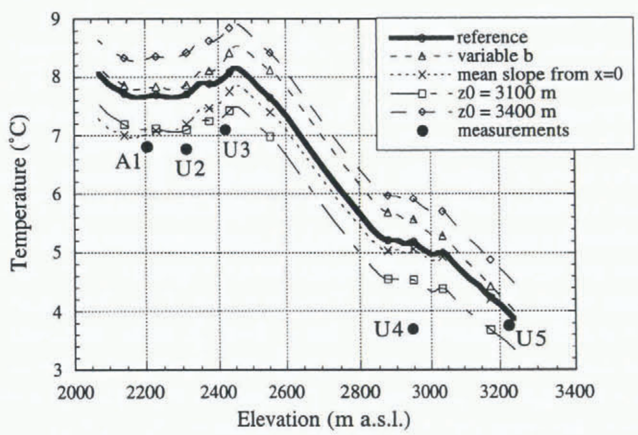

Fig. 8. Sensitivity of the calculated temperature distribution to variations in b and z_0. In all calculations, b was computed from the mean slope between $x = -x_0$ and the terminus, except in a calculation which takes into account variations in slope along the glacier ("variable b"), and a calculation with a value of b computed from the mean slope between $x = 0$ and the terminus. In all calculations, z_0 was taken as the elevation at the top of the flowline (3240 m a.s.l.), except in two experiments with $z_0 = 3100$ and 3400 m a.s.l. The measurements are mean values for approximately 48 days.

A question that may arise is to what extent the results of the foregoing analysis depend on the choice of climate station. We used the station on Sonnblick because of its proximity to the Pasterze and because it has a time series of atmospheric temperature of considerable length (from 1887 on), which has been studied and described in detail (see Auer and others, 1992). Furthermore, its elevation is almost the same as the top of the PASTEX flowline. In order to investigate the influence of the choice of climate station, the parameters L_R and x_0 were also determined by means of data from seven other climate stations located within 40 km of the Pasterze. The corresponding climate sensitivities along the glacier are shown in Figure 9. Along the entire glacier the climate sensitivity is somewhat greater when the data used are from stations located at lower (1010–1242 m a.s.l.) instead of higher elevations (1973–2304 m a.s.l.). In fact, L_R is somewhat larger (10 730–11 610 m) for data from the stations at lower elevations, and smaller (8840–9150 m) for data from the stations at higher elevations, except for Bad Gastein (9020 m), and x_0 is smaller (870–1350 m) for the stations at lower elevations, and larger (2260–3080 m) for the stations at higher elevations. Sonnblick is the only summit station and, in terms of the climate sensitivity, occupies an intermediate position between the two groups of stations. The reasons for these effects are unknown, but it can be concluded that the results of the analysis presented above are not due to the specific position of the Sonnblick observatory, on the summit of a mountain.

Finally in this section, we discuss data from three other glacio-meteorological experiments. These were carried out on Rhonegletscher in Switzerland (Funk, 1985), Storglaciären in Sweden (Hock and Holmgren, 1996) and Haut Glacier d'Arolla in Switzerland. Some specifications are given in Table 1. Datasets from two or three climate stations were available for all of these glacio-meteorological experiments. Among these, the datasets having the highest correlation with the data obtained on the glacier were selected. Data from climate stations located in the vicinity of glaciers, like those from Tarfala (near Storglaciären; see Grudd and Schneider, 1996), were discarded because glaciers may have influenced the measured 2 m temperatures. The calculated values of $\mathrm{d}T/\mathrm{d}T_{cs}$ are shown in Figure 5, together with those from the Pasterze. The values from Rhonegletscher

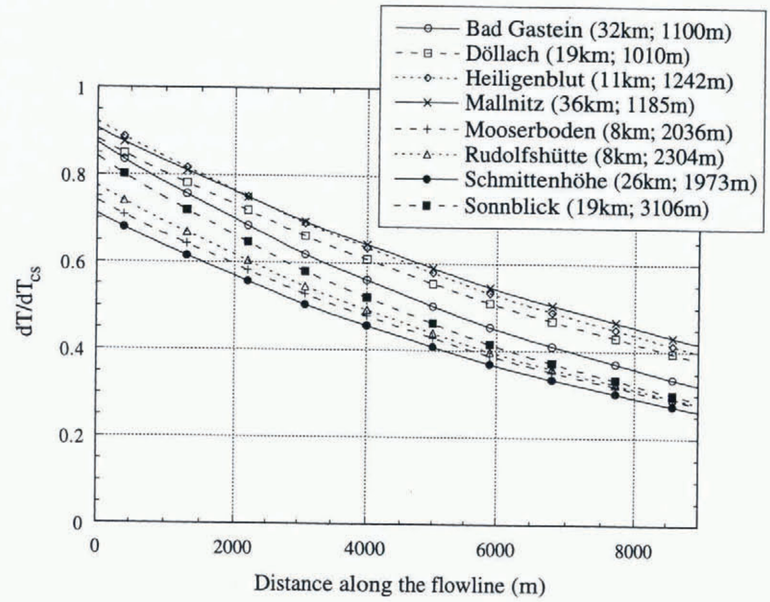

Fig. 9. Variation in the climate sensitivity (sensitivity of the local temperature to the temperature at a climate station) along the glacier, derived from data from eight climate stations. Curves gives best fits of Equation (12) to the climate sensitivities at the individual stations.

and the lower meteorological station (LMS) on Haut Glacier d'Arolla are compatible with those from the Pasterze, but the values from the upper meteorological station (UMS) on Haut Glacier d'Arolla and the station on Storglaciären are relatively high, with the largest deviation for Storglaciären. This may be caused by lower frequencies of the glacier wind at the latter two sites. An indication of the frequency of the glacier wind is provided by the directional constancy, which was equal to 0.35 during the experiment at LMS and 0.20 at UMS. These values are much lower than the directional constancies found during PASTEX (see Fig. 1). Thus, at UMS, where according to the directional constancy the glacier wind seems to occur less frequently than at LMS, the deviation of $\mathrm{d}T/\mathrm{d}T_{cs}$ from the values found on the Pasterze is indeed larger than at LMS. Wind data are not available for the experiments performed on Rhonegletscher and Storglaciären. Strong gradient winds may be the cause of a lower frequency of the glacier wind on Storglaciären than on the other glaciers and therefore of a relatively high value of $\mathrm{d}T/\mathrm{d}T_{cs}$.

5. IMPLICATIONS FOR THE SENSITIVITY OF THE MASS BALANCE TO CLIMATIC CHANGE

In this section we investigate the effect of atmospheric warming on the surface mass balance of glaciers, using the glacier-wind model proposed in the previous sections. In this way we can link the 2 m temperature outside the thermal regime of the glacier to the 2 m temperature above the glacier. The latter is then related to the mass balance by means of a surface energy-balance model that is forced by five meteorological variables: namely, atmospheric temperature (2 m), amount of precipitation, wind speed (2 m), cloud amount and atmospheric humidity (2 m). The models are applied to the Pasterze, and the daily means of meteorological variables measured on Sonnblick are used as input.

These input variables are converted to local input variables along the glacier in the following way:

(a) In order to compute the 2 m temperature distribution along the glacier, one first has to determine the elevation at which the temperature is equal to the melting point. This is done by assuming a constant lapse rate of $-0.007\ \mathrm{K\,m^{-1}}$, which is the average lapse rate above the glacier-wind layer established during PASTEX by means of balloon soundings. This lapse rate is then also used to compute the temperature distribution above the elevation where the temperature is 0°C. At lower elevations, the glacier-wind model is applied, setting $x_0 = 0$ whenever the elevation at which the temperature is 0°C is below the top of the flowline (3240 m a.s.l.). To obtain a better agreement between calculated and measured temperatures for PASTEX, we made a correction of −0.74°C to

Table 1. Specifications of the glacier and climate stations that provided data for the determination of $\mathrm{d}T/\mathrm{d}T_{cs}$ for several glaciers. In the analysis, daily mean temperatures are used

Glacier, station	*Elevation*	*Distance from top of flowline*	*Number of samples*	*Climate station*	*Elevation*	*Distance from glacier station*
	m a.s.l.	km			m a.s.l.	km
Rhonegletscher	2840	3.5	59	Grimsel Hospiz	1950	7
Storglaciären, 9B	1370	2.0	72	Nikkaluokta	470	19
Haut Glacier d'Arolla, UMS	2900	1.95	33	Grand St Bernard	2472	30
Haut Glacier d'Arolla, LMS	2725	3.2	38	Grand St Bernard	2472	30

the computed temperatures (see Fig. 7), both above and below the elevation at which the temperature is 0°C.

(b) The local amount of precipitation is the product of a factor and the amount of precipitation measured on Sonnblick. This factor is computed as the inverse tangent of the distance along the flowline (see Fig. 10). It is equal to 1.3 at the top of the flowline, whereas at $x = 17$ km (Heiligenblut) it yields a mean annual precipitation equal to the locally measured amount of 1200 mm a^{-1} (Tollner, 1952, fig. 19). While keeping these two conditions fixed, we varied the parameters that determine the shape of the inverse tangent in order to tune the model in terms of the mass balance (see below). In the calculations, the rain/snow threshold is set at a free-atmosphere temperature of 2°C. The local free-atmosphere temperature is computed from the temperature on Sonnblick on the assumption of the above-mentioned constant lapse rate (-0.007 K m^{-1}).

(c) The PASTEX measurements did not reveal any relationship between the 2 m wind speed above the glacier, on the one hand, and the wind vector on Sonnblick or the local 2 m temperature, on the other hand. Therefore, in the model, daily values of the wind speed were not related to any other model variable but were generated at random by the computer in such a way that the wind-speed frequency distribution recorded at Al was reproduced (see Greuell and others, in press).

(d) Cloud amount is calculated as 1 minus fractional sunshine duration on Sonnblick, and taken as a constant along the glacier. Here fractional sunshine duration is defined as measured sunshine duration divided by potential sunshine duration.

(e) Constancy along the glacier is also assumed for the 2 m humidity mixing ratio, which is set equal to 95% of the value measured on Sonnblick, as suggested by the PASTEX data.

The energy-balance model used here has been applied previously to the Greenland ice sheet. Details are described in Greuell and Konzelmann (1994). Here, only an outline of the model is given. The model computes the mass balance at a specific point and is forced by the above-mentioned meteorological variables that describe local conditions. Spatial variations in mass balance can be studied by repeating the calculations at a number of points along the glacier.

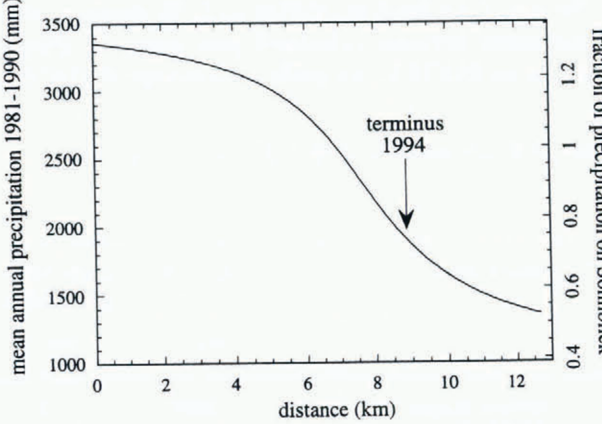

Fig. 10. In the model this inverse tangent is used to compute the local amount of precipitation as a fraction of the amount of precipitation measured on Sonnblick.

In this case, computations were made at 11 points along the Pasterze spaced at elevation intervals of 115 m and ranging in elevation between 2075 and 3225 m a.s.l.

The model consists of two parts.

(a) *Computation of the fluxes between atmosphere and glacier.*

(1) The local input variables (cloud amount and 2 m temperature and humidity) are inserted into parameterizations derived from the PASTEX data (see Greuell and others, in press) to compute the incoming fluxes of short-wave and long-wave radiation.

(2) Another parameterization gives the albedo as a function of surface conditions. The ice albedo is 0.2, whereas the snow albedo decreases exponentially from 0.8 to 0.5 with the total amount of time that the actual surface material has been wet since its deposition (called "wet-age" from now on). The e-folding time is 24 days. This parameterization was tuned by means of the PASTEX data.

(3) Outgoing long-wave radiation is calculated from the surface temperature. The emissivity is assumed to be equal to 1.0.

(4) The turbulent fluxes are obtained from the 2 m and surface values of temperature, humidity and wind speed, the Businger–Dyer flux-profile relationships (see Stull, 1988) and roughness lengths for wind speed (z_{0m}), temperature (z_{0T}) and water vapour (z_{0q}). Values of z_{0m} for ice (4.4 mm) and wet snow (2.3 mm) were derived from the PASTEX data, whereas the value for dry snow (0.12 mm) was found by Ambach (1977) on the Greenland ice sheet. The equations proposed by Andreas (1987) were used to calculate z_{0T} and z_{0q} from z_{0m} and the friction velocity.

(b) *Computation of processes within the glacier.* Temperature, density, water content and wet-age are computed on a one-dimensional grid extending approximately 25 m downwards into the glacier. The following processes are described on this grid: formation; percolation; refreezing and runoff of meltwater; conduction; penetration of short-wave radiation; and densification by compaction. As a result the formation of slush and superimposed ice can be simulated. The energy exchange with the atmosphere constitutes a boundary condition.

The experiments were performed with a time-step of 3 minutes. Although input data consist of daily means measured on Sonnblick, the daily variation is simulated by computing the zenith angle of the sun and by imposing at Sonnblick a sinusoidal diurnal temperature variation with an amplitude of 2.2°C.

Simulations were performed for 1980–89 because for this period the mass-balance output can be compared to direct stake measurements. The advantage of forcing by daily means over a longer time period (9 years) instead of forcing by climatological means is that the effect of variations around the climatological mean are taken into account. This is probably important; for example, snowfall events during the ablation season have a double impact on the mass balance: not only do they increase the accumulation, they also increase the albedo and therefore reduce ablation (see e.g. Greuell and Oerlemans, 1986).

First a reference run (run 1) was made in accordance with the above-mentioned specifications. The mean simulated mass-balance profile for the period 8 September 1980–19 September 1989 is plotted in Figure 11 and compared to

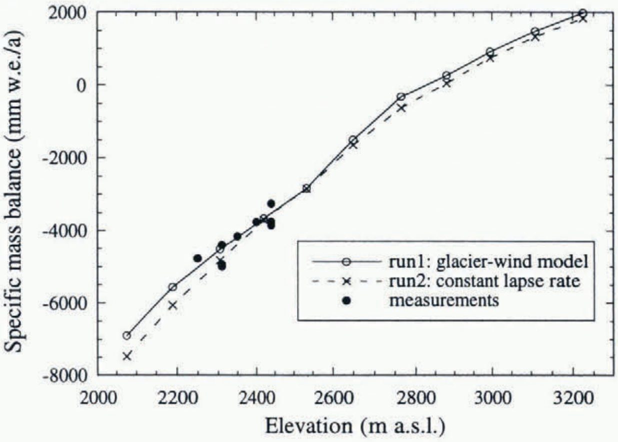

Fig. 11. Mean specific mass balance for the Pasterze, 8 September 1980–19 September 1989, against elevation as measured (see Tintor, 1991) and computed with a surface energy-balance model. Different curves are obtained with different methods to compute the 2 m temperature along the glacier.

measurements for the same period made on the tongue (from Tintor, 1991). Mass-balance data are unavailable for the upper part of the glacier. It can be seen that measurements made at the same elevation, but displaced relative to each other in the lateral direction of the glacier, differ by up to 600 mm w.e. a^{-1}. Sensitivity experiments performed with the energy-balance model indicate that the most likely causes of differences of this order of magnitude are albedo variations in the lateral direction, and not differences in shading by the surrounding slopes. As already mentioned, the tuning procedure consisted of adjusting the inverse tangent describing total precipitation as a function of distance along the glacier until an optimal agreement was reached between calculated and measured mass balance according to the least-squares criterion.

In a second run, the 2 m temperature along the glacier was computed on the assumption of a constant lapse rate of $-0.007\ \mathrm{K\,m^{-1}}$ under all conditions. The resulting mass-balance profile is also shown in Figure 11. Both the glacier-wind model and a constant lapse rate are found to yield a specific mass balance which is almost linear with elevation. However, when compared to the assumption of a constant lapse rate, the glacier-wind model leads to somewhat larger mass-balance gradients in the steeper parts of the glacier (2550–2850 m a.s.l.) and somewhat smaller mass-balance gradients in flatter parts of the glacier (2200–2450 and 2850–3050 m a.s.l.). The difference can be attributed to the temperature distribution during melt conditions (see Fig. 6). It should be pointed out here that during PASTEX the increase in the 2 m temperature with elevation above the glacier tongue led to a reversed mass-balance gradient, i.e. ablation increased with elevation (see Greuell and others, 1995).

Next, several runs were made to study the effect of 1 °C atmospheric warming at the climate station (Sonnblick). In all runs the humidity input was adapted on the assumption that the relative humidity above the glacier is not affected by the temperature change. Thus, specific humidity increased. In total, five runs are discussed, all based on the reference run, which implies that the glacier-wind model was used to compute undisturbed 2 m temperatures. Temperature perturbations along the glacier were also computed with the glacier-wind model or by assuming a constant lapse rate, as specified below. Changes in specific mass balance relative to the reference experiment are depicted in Figure 12 as a function of elevation, and average changes over the 11 elevations are listed in Table 2.

During run 3, local temperatures were increased by 1 °C, but the amount of solid precipitation and the albedo were the same as in the reference run, so that changes in accumulation and the albedo feedback were neglected. Since on the assumption of a constant lapse rate the temperature increase along the glacier is equal to that at the climate station (1 °C in this case), this experiment computes the climate sensitivity as if the lapse rate were constant. Figure 12 shows that the change in specific mass balance increases with decreasing elevation. This can be ascribed to the lengthening of the ablation season with decreasing elevation. The relation between change in specific mass balance and elevation is almost linear.

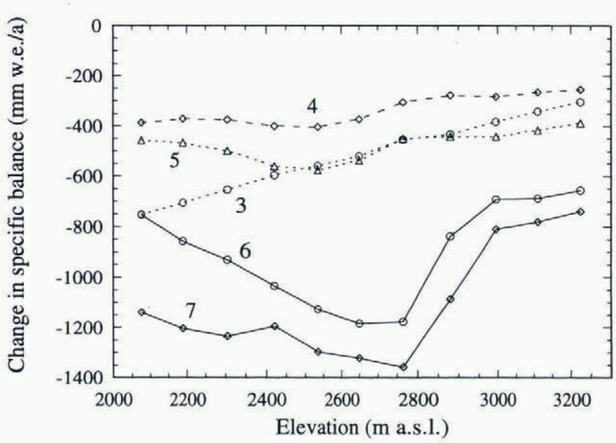

Fig. 12. Change in specific mass balance at the Pasterze, 8 September 1980–19 September 1989, after 1 °C atmospheric warming at Sonnblick, against elevation as computed with a surface energy-balance model.

- *run 3: temperature change = 1 °C along the entire glacier; no changes in amount of solid precipitation and albedo.*
- *run 4: temperature change computed with the glacier-wind model; no changes in amount of solid precipitation and albedo.*
- *run 5: temperature change computed with the glacier-wind model; change in amount of solid precipitation computed; no change in albedo.*
- *run 6: temperature change computed with the glacier-wind model; changes in amount of solid precipitation and albedo computed.*
- *run 7: temperature change = 1 °C along the entire glacier; changes in amount of solid precipitation and albedo compared.*

Table 2. Perturbation of specific mass balance (db) on the Pasterze after 1 °C atmospheric warming as computed with a surface energy-balance model. The table shows the average for 11 elevations. Settings during the different model runs are further explained in the text

	db mm w.e. a^{-1}
run 3: constant lapse rate; no change in solid precipitation and albedo	−520
run 4: glacier-wind model; no change in solid precipitation and albedo	−340
run 5: glacier-wind model; no change in albedo	−470
run 6: glacier-wind model	−900
run 7: constant lapse rate	−1110

During run 4, settings were the same as during run 3. However, in run 4 we used the glacier-wind model to compute the 2 m temperature distribution after imposing the temperature perturbation at Sonnblick. Therefore, during melt conditions local temperature changes are smaller than 1°C, and the change in mass balance is less than in the previous run. The reduction increases with the distance along the glacier, as can be understood from Equation (12). In fact, this experiment yields the change in ablation without inclusion of the albedo feedback.

Run 5 was as run 4. However, the amount of solid precipitation was not the same as in the reference run, but was computed with the normal procedure, that is from the amount of total precipitation and the perturbed free-atmosphere temperature. Therefore, after atmospheric warming, the threshold temperature between solid precipitation and rain is more often exceeded than before, which results in a reduction in the accumulation. The effect increases with elevation, because the amount of total precipitation increases with elevation. This experiment yields the total change in specific mass balance (ablation plus accumulation) without inclusion of the albedo feedback.

In order to include the albedo feedback, during run 6 the albedo was generated internally. This experiment represents the best estimate of the change in specific mass balance after 1°C atmospheric warming. As already demonstrated by Oerlemans and Hoogendoorn (1989), the albedo feedback is important. It causes the change in specific mass balance to decrease from −470 mm w.e. a^{-1} (run 5) to −900 mm w.e. a^{-1} (run 6), i.e. by 91%. The magnitude of the albedo feedback is a complicated function of elevation. It reaches its maximum at medium elevation where the increase in the length of the period with ice exposed at the surface reaches its maximum. At higher elevations, snow is exposed at the surface throughout the year, both with and without atmospheric warming, so only increased wet-ages of the snow influence the albedo. On the other hand, at very low elevations, which are not considered in the calculations, because there is no glacier now, accumulation would be zero, so ice would be at the surface there throughout the year. This implies that there would be no albedo feedback. This behaviour is approached asymptotically when elevation decreases, which explains the weakness of the albedo feedback in the lower part of the glacier.

Finally, during run 7, settings were the same as during run 6, but the temperature perturbation was set at +1°C along the entire glacier. The difference between runs 7 and 6 in Figure 12 is due to the fact that the sensitivity of the mass balance to 1°C warming was overestimated because of the assumption of a constant lapse rate. On average along the glacier, the difference is equal to −210 mm w.e. a^{-1} (see Table 2), that is 23% of the sensitivity computed with the glacier-wind model.

Ice-flow models that are used to relate length fluctuations of glaciers to climatic variations have to incorporate an assumption about the relation between a climatically induced perturbation in the mass balance and elevation. Usually it is assumed that the annual perturbation from a reference profile is constant with elevation (see e.g. Greuell, 1992; Oerlemans, 1997). This is in line with measurements that show that on alpine glaciers year-to-year variations in specific mass balance are nearly independent of elevation (Kuhn, 1984). The experiments presented here suggest that without albedo feedback the relation is more or less linear (run 5, Fig. 12), but the relation becomes much more complicated if the albedo feedback is taken into account (run 6, Fig. 12). However, it is dangerous to use the relation between the mass-balance perturbation and elevation depicted in Figure 12 in the ice-flow models, because it is not yet known whether this relation is universally valid. Therefore, the assumption that the perturbation is constant with elevation seems to be the best option at the moment.

6. CONCLUSIONS AND DISCUSSION

This paper presents a simple thermodynamic glacier-wind model that can be used to compute the 2 m temperature distribution along a melting glacier. The model takes only two processes into account, adiabatic heating of air flowing down the glacier, and cooling due to exchange of sensible heat with the underlying surface. In order to find an analytical solution for the temperature distribution, one needs to assume a constant slope. However, if the solution is first written in terms of potential temperature as a function of the distance along the glacier on the basis of the mean slope, and is then transformed into temperature by means of the surface profile of the glacier, one can find a solution for a glacier with a variable slope. This solution seems to be satisfactory for the Pasterze (Austria), as shown by comparison with a numerical solution that takes variations in slope into account, and by comparison with data. Model solution and measurements demonstrate that the relation between potential temperature and the distance along the glacier is more linear than the relationship between temperature and elevation. Note that a linear relationship between temperature and elevation (a constant lapse rate) has been assumed in many studies.

The use of the glacier-wind model instead of calculations based on the assumption of a constant lapse rate leads to a considerable difference in the predicted sensitivity of the local temperature to changes in temperature outside the thermal regime of the glacier. According to the glacier-wind model, this sensitivity decreases exponentially from 1 at the top to 0 at very large distances from the top, but if a constant lapse rate is assumed it is 1 along the entire glacier. Data for the Pasterze and some other glaciers confirm the outcome of the glacier-wind model.

This difference in the predicted sensitivity of the 2 m temperature has implications for the sensitivity of the mass balance to atmospheric warming. A surface energy-balance model forced by the 2 m temperature outside the thermal regime of the glacier was used to quantify this sensitivity for the Pasterze. The model was tuned with the precipitation. A match between measured and calculated mass balance could be obtained when reasonable functions were taken to describe the largely unknown variation of precipitation with the distance along the glacier. This was possible both when the glacier-wind model was used and when a constant lapse rate was assumed to compute 2 m temperatures. Therefore, such a tuning procedure does not provide extra support for the correctness of the glacier-wind model. However, the difference between the two methods in terms of the sensitivity of the mass balance to atmospheric warming is 23%. Atmospheric warming of 1°C outside the thermal regime of the glacier led to a change of −900 mm w.e. a^{-1} in the specific mass balance when the perturbation of the 2 m temperature along the glacier was obtained with the glacier-wind model. On the assumption of a constant lapse rate, however, the change was −1110 mm w.e. a^{-1}. This means that the estimates of the effect of atmospheric warming on the mass balance made by, for example, Oerlemans and Hoogendoorn (1989), Oerlemans and Fortuin (1992) and Jóhannesson and others (1995) are too high.

The glacier-wind model, however, does have some limitations. The model can only be used if there is a glacier wind. Thus, minimum requirements for application are a melting surface and temperatures higher than 0°C just above the surface. Even if these requirements are fulfilled, a glacier wind does not necessarily develop. Strong gradient or mountain and valley winds may have a considerable influence on the near-surface wind regime. This was possibly the case during the PASTEX experiment in the upper part of the Pasterze, where the measured temperature distribution could not be simulated with the model, and during an experiment on Storglaciären, where the sensitivity of the local temperature to changes in temperature outside the thermal regime of the glacier turned out to be incompatible with that on the other glaciers considered in this paper.

Also, analysis of data from the marginal zone of the Greenland ice sheet (Søndre Strømfjord transect; not shown here) showed that at that location the "climate sensitivity" of the 2 m temperature increased towards the margin, i.e. it increased with x. Therefore, the model is not applicable there, although the katabatic wind dominates the near-surface wind regime (see e.g. Oerlemans and Vugts, 1993). This may be understood in terms of the heat budget. The two-term balance in the heat budget assumed in the glacier-wind model presented here was found for the Pasterze (Van den Broeke, in press), but above the Søndre Strømfjord transect the budget becomes more complicated because divergence of long-wave radiation is also important (Van den Broeke and others, 1994). This difference in the heat budgets of the Pasterze and the Greenland ice sheet is probably due to the difference in length and/or height scale of the katabatic layer.

In summary, according to its theoretical basis, the glacier-wind model can be applied to melting glaciers, provided gradient and mountain and valley winds do not perturb the glacier-wind layer. The available data from four mid-latitude glaciers suggest that the model is indeed applicable to three of these glaciers. Thus, over most melting, mid-latitude glaciers, disrupture of the glacier-wind layer by other winds is probably of lesser importance. According to the data, the model cannot be used to compute the temperature distribution above the ablation zone of the Greenland ice sheet.

How can the glacier-wind model presented here be applied to other glaciers? If sufficient temperature data are available, the values of the unknown parameters x_0, L_R, b, z_0 and γ can be determined as in this study, but what is to be done if no or insufficient data are available? According to Equation (5), the response length (L_R) is proportional to the thickness of the glacier-wind layer (H). The glacier-wind model of Ohata (1989) predicts that there will be an almost linear relationship between H and the distance along the glacier, but the model presented here successfully reproduces the measured temperature distribution over the Pasterze on the assumption of a constant H. Van den Broeke (in press) also concluded that H is nearly constant, after analyzing the heat, momentum and moisture budget of the katabatic layer over the tongue of the Pasterze. If H is in fact constant and if it is assumed that the local thickness of the glacier-wind layer is not affected by processes occurring down-glacier, H should be independent of glacier length. Therefore, it seems reasonable to take L_R on other glaciers equal to the value found for the Pasterze.

The distance x_0 possibly depends on the geometry of the upper part of the glacier and should therefore have different values for different glaciers. However, as long as the nature of this dependence is not determined, x_0 remains an empirical constant. In the absence of data it is advisable to take the same value as found for the Pasterze. In any case, the influence of the value of x_0 is slight at distances along the glacier much larger than x_0. Note that Figure 5 suggests that L_R and x_0 on other glaciers are indeed similar to the values found for the Pasterze.

As far as the constant b is concerned, the present study suggests that it should be computed from the mean slope between $-x_0$ (elevation z_0) and the terminus of the glacier. The height z_0 might be used as a tuning parameter, as shown in Figure 8. However, alternatively the model could be tuned by applying a constant correction to the temperatures found. In that case, or in the absence of temperature data, z_0 might be taken equal to the elevation at the top of the flowline. Finally, γ could be taken as the mean lapse rate in the free atmosphere, as inferred from local upper-air soundings.

Finally, this paper demonstrates (Fig. 9) that L_R and x_0 do not depend very much on the elevation or the type of location (valley or summit) of the climate station considered. This conclusion has some bearing on predictions of future melt, which are based on predictions of changes in atmospheric temperature produced by global atmospheric models (see e.g. de Wolde and others, 1996). Given that the size of the gridboxes of these models is much larger than the typical wavelength of mountains and valleys, one might well wonder at what elevation and at which type of location the predicted temperature changes are valid in mountainous areas. Whatever the answers to these questions are, the calculated change in the 2 m temperature along the glacier is more or less the same, owing to the small variation in the L_R's and x_0's occurring with different topographic settings of the climate station.

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