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Glacier winds and parameterisation of the related surface heat fluxes

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ABSTRACT

The katabatic flow over glaciers is studied with data from automatic weather stations (AWS). We analyse data from the Morteratschgletscher (Switzerland), Vatnajökull (Iceland) and West Greenland, and conclude that katabatic flow is very common over melting glacier surfaces and rarely disrupted by the large-scale flow. Over small and medium-size glaciers the height of the wind maximum is generally low (typically 10 m), and vertical temperature differences near the surface are very large (up to 15 K over 4 m). In glacier mass-balance models there is a great need for parameterisations of the surface heat flux. We develop a simple method to estimate the sensible heat flux F_h associated with the glacier wind. It is based on the classical Prandtl model for slope flows. We set the turbulent exchange coefficient proportional to the maximum wind speed (velocity scale) and the height of the wind maximum (length scale). The resulting theory shows that F_h increases quadratically with the temperature difference between the surface and the ambient atmosphere; F_h decreases with the square root of the potential temperature gradient of the ambient atmosphere; and F_h is independent of the surface slope.

1. Introduction

It has been known for a long time that on fair summer days shallow downslope winds develop over melting glaciers (e.g. Tollner, 1931; Hoinkes, 1954). In the early days of dynamic meteorology it was also recognised that these glacier winds can be described as a special case of slope winds (Defant, 1949). Glacier winds are driven by the generation of negative buoyancy close to the surface, just like nocturnal drainage flows. In the last 20 years a number of studies have been devoted to nocturnal drainage flows (e.g. Yamada, 1983; Horst and Duran, 1988), and very few to the glacier wind.

Katabatic flows over ice sheets and glaciers occur on a wide range of scales. The katabatic flows in the margin of the Antarctic Ice Sheet are well known for their strengh and persistence. Typical spatial scales are a few hundred of metres in the vertical and hundreds of kilometres in the horizontal. As a consequence, the Coriolis force is an important term in the momentum budget (Ball, 1956; Parish and Waight, 1987; Van den Broeke et al., 1994). At the other end of the spectrum is the katabatic flow over small glaciers. In this case the Coriolis force is one or two orders of magnitude smaller than the buoyancy force, and the depth of the katabatic layer is only 10–50 m (Van den Broeke, 1997). In this paper the focus is on the boundary layer over small and medium-size glaciers with significant ablation zones. We use the term 'glacier wind' to describe the katabatic flow developing over a surface that is at the melting point for most of the time.

The glacier wind acts as a heat pump for the glacier surface. In spite of the extremely stable stratification (up to 5 K m^{-1} in the lower metres), turbulence is generated that brings heat to the surface. Interestingly, a larger heat flux implies a larger buoyancy force, and the wind speed will increase. In the end friction will limit the strength of the flow, of course, but, given

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the strong stratification, appreciable wind speeds are maintained (typically 6 m s^{-1}). Although meteorological experiments on glaciers have shown that the largest contribution to the melt energy comes from solar radiation (Greuell et al. 1997; Oerlemans et al., 1999), the *sensitivity* of the melt rate to temperature change is determined by changes in the longwave balance and in the turbulent heat flux (e.g. Oerlemans, 2001). Therefore, to quantify the sensitivity of glacier mass balance to temperature change we need to know how the turbulent exchange is related to the conditions in the ambient atmosphere. Clearly, this cannot be achieved without a thorough understanding of the dynamics of the glacier wind. This topic has received little attention, except for an early attempt by Kuhn (1978). Kuhn tried to relate the sensible heat flux at the surface to the height of the wind maximum but had very few data to pursue and check the idea.

Apart from the significance with respect to climate change, the glacier wind is also interesting from a purely meteorological point of view. Over glacier tongues the surface temperature remains at the melting point for long periods of time (sometimes for several months), providing a lower boundary condition for temperature and vapour pressure that is much simpler than for nocturnal drainage flows. Moreover, the flow is generally stronger with a well defined wind maximum, and the terrain is more homogeneous (a constant slope over larger areas). The drawbacks to obtaining high-quality data, however, are the poor accessibility and the unstable ice surface. (Melt rates between 5 and 10 m in a summer make it difficult to establish stable structures for mounting instruments.)

The micrometeorological processes that determine glacier melt have been identified in a number of meteorological experiments (e.g. Ambach, 1963; Björnsson, 1972; Wendler and Weller, 1974; Munro and Davies, 1978; Hogg et al., 1982; Munro, 1989; Ohata et al., 1989; Ishikawa et al., 1992). However, most of these studies were restricted to a single location and/or were of short duration. During the last decade more extensive meteorological experiments have been performed in which five or more stations were operated simultaneously for periods of a few months in summer (Oerlemans and Vugts, 1993; Greuell et al., 1994; Oerlemans et al., 1999). An overview of these experiments can be found in Oerlemans (1998). Meanwhile the development has continued, and the Institute for Marine and Atmospheric Research (Utrecht University) is now operating quasi-permanent weather stations (Fig. 1) in the melt zones of a number of glaciers (central West Greenland; Vatnajökull, Iceland; Morteratschgletscher, Switzerland; Hardangerjökulen, Norway; Storbreen, Norway). Some of these stations have produced records several years long.

Fig. 1. A picture of an automatic weather station on a melting glacier surface (Hardangerjökulen, Norway; photograph taken by Wim Boot in August 2000). The construction with the meteorological sensors (middle of the picture) is about 6 m tall and stands freely on the surface. It sinks with the melting ice surface. The tripod (left) is drilled into the ice. A sonic ranger is mounted on it to measure ice melt and snow accumulation. In winter, when snow covers the tripod, a second sonic ranger in the top of the mast (below the wind sensor) takes over.

In this paper data from automatic weather stations on different glaciers are used to characterise katabatic flow over glaciers. After a general overview, we consider the vertical structure of the glacier wind, based on measurements carried out with a profile mast and a cable balloon on the Pasterze glacier, Austria, in the summer of 1994. Finally the theory of katabatic flow is briefly reviewed and a method is presented to estimate the surface heat flux from a simple theory.

2. A look at data from automatic weather stations

We first consider scatter plots of wind direction, wind speed and air temperature for three AWS on different glaciers. The AWS on the Morteratschgletscher is located on the lowest part of the glacier, where the annual melt of ice is about 6 m. The glacier is about 7 km long and spans an altitudinal range of 2000–4000 m above sea level. It flows in a northerly direction. The altitude of the AWS is 2105 m and the local slope is about 5°. Air temperature, humidity, pressure, windspeed, wind direction, and incoming and outgoing longwave and shortwave radiation are sampled every 2 min and half-hourly mean values are logged. In winter snow temperatures are also measured. The station stands freely on the ice and sinks with the melting glacier surface. This keeps the height of the sensors approximately constant until snow falls. The height of the sensors is 3.5 m minus the snow depth. The snow depth and ice melt are measured with a sonic ranger mounted to a structure that is drilled in the ice.

We now discuss the relation between windspeed, wind direction and air temperature as summarised for some stations in Fig. 2. The plot of wind direction against windspeed for the Morteratschgletscher contains all 30-min values for the year 1998 (i.e. 17 520 data points). It is clear that the wind almost always blows downglacier. More detailed analysis, not illustrated here, shows that on this glacier tongue the occurrence of katabatic flow has no clear seasonal signal.

In the classical description of katabatic flow, the forcing is determined by the temperature deficit in the near-surface layer. This deficit is defined as the difference between the actual air temperature T_a and the temperature of the ambient atmosphere, extrapolated to the same altitude. Generally speaking, T_a measured at a few metres height above the glacier surface contains little information about this deficit because the ambient temperature is not known. However, when the surface is melting and T_a is above the melting point (T_m) the temperature deficit must be negative and katabatic flow must be generated. The buoyancy forcing is proportional to the temperature deficit, and this is reflected in Fig. 2. For $T_a < T_m$ there is little correlation between temperature and windspeed, but for $T_a > T_m$ there is a clear tendency for stronger winds when the temperature is higher. In fact, cases with high temperature and low windspeed are very rare.

On the Morteratschgletscher the glacier wind exhibits a great regularity. Figure 3 shows records of air temperature and wind direction and speed for an 8-d period of fine weather in October 1995. The wind direction hardly varies (it is downglacier all the time) whereas the windspeed shows a marked daily cycle. The windspeed slightly lags air temperature. Air temperature as measured at the mountain weather station Corvatsch (3297 m, 9 km distance to the AWS) and at Samedan (1704 m, 14 km distance to the AWS) are also shown. Note the very large daily amplitude at Samedan, which is located in a wide flat valley. The daily amplitude at the mountain station is rather small.

In the middle panel of Fig. 2 data are shown for an AWS on one of the outlet glaciers (Breidamerkurjökull) of Vatnajökull, the largest ice cap of Iceland. Despite the larger spatial scale and the fact that the glacier is in the middle of the North Atlantic storm track bringing strong large-scale winds, the flow close to the ice surface is still mainly of a katabatic origin. The detailed meteorological experiment in the summer of 1996 (Oerlemans et al., 1999), during which a large number of weather stations were placed across Vatnajökull, has clearly showed that the preference of wind direction along the local fall line disappears gradually when going upglacier. This is related to the fact that the surface slope becomes smaller and, as a consequence, the magnitude of the buoyancy forcing becomes comparable to the synoptic-scale pressure gradient.

The relation between windspeed and air temperature is somewhat different now. For temperatures above the melting point there is a slight tendency for higher wind speeds to occur with higher temperature, but the relation is quite weak (much weaker than seen for AWS on the Morteratschgletscher and in central West Greenland, see below).

The third AWS from which data are shown in Fig. 2 is operated in the melt zone of the Greenland Ice Sheet. The station is located at a distance of 6 km from the ice edge, at an altitude of about 500 m. On the local scale (10 m) the terrain is quite rough and bumpy. The

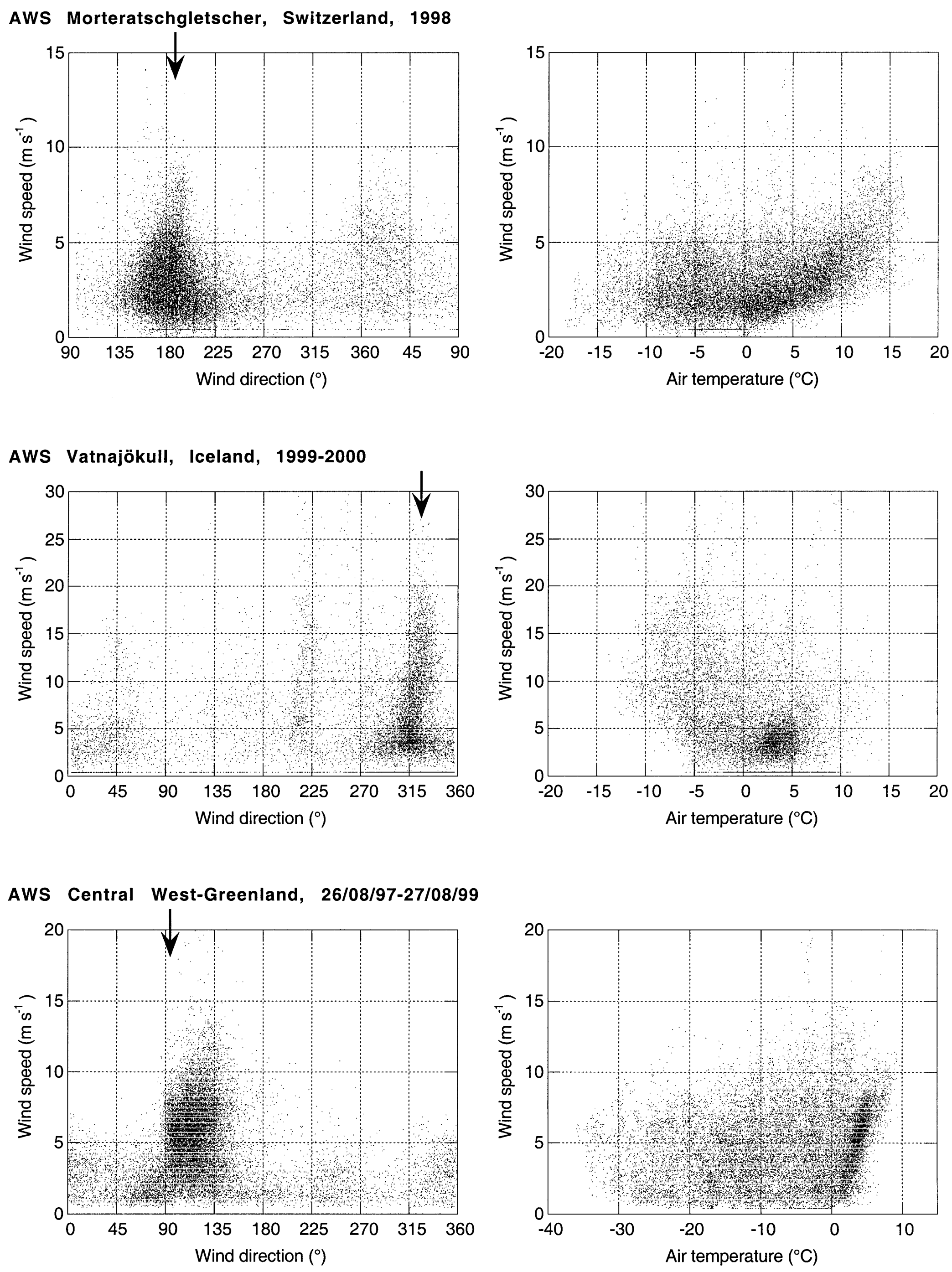

Fig. 2. Scatter plots from automatic weather stations on glaciers, showing the relation between wind direction, wind velocity and air temperature. For the Morteratschgletscher each data point represent a 30-min value, for the other glaciers a 60-min value. The number of data points in each plot is about the same (~17 500). The arrows indicate the downslope wind direction.

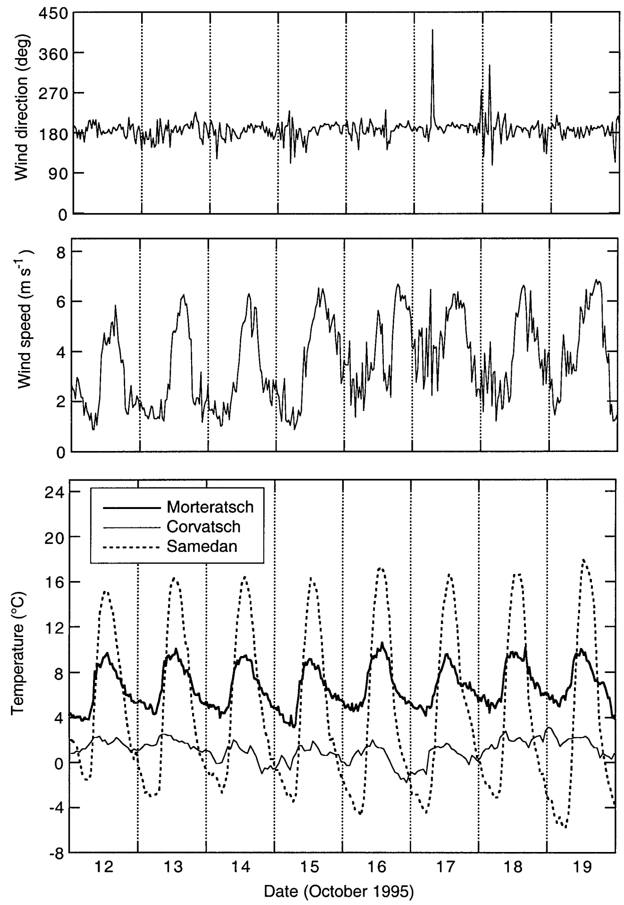

Fig. 3. Daily cycles of temperature, wind speed and wind direction at the Morteratsch AWS, compared with temperature at Samedan (1705 m) and Corvatsch (3297 m).

surface slope on the larger scale (1 km, say) is about 3°. The region is very dry and the snow depth in winter is small. It hardly affects the sensor height, which is about 3.5 m. The data from this AWS differ in at least one aspect from the data of the other stations: although there is a well defined preferred direction of the flow, this direction is turned away about 20° with respect to the fall line. Apparently, the larger time and spatial scale, in a combination with the smaller surface slope, makes the Coriolis acceleration a significant term in the momentum balance. The relation between temperature and wind speed for $T_a > T_m$, as found for the other glaciers, shows up again. In fact, this relation appears to be quite strong.

The AWS data discussed above suggest that over the melting zones of glaciers katabatic flow is the rule. Therefore, parameterisation of the sensible heat flux at the glacier surface should include the effects of katabatic flow in one way or another. We will come back to this point in section 5.

3. Vertical structure of the glacier wind

The glacier wind is characterised by a distinct wind maximum. To observe the structure of the jet in detail a well equipped meteorological tower of typically 50 m height would be the best tool. However, well developed glacier winds occur over ablation regions where the melt rates are high (typically between 5 and 10 cm of ice per day). In combination with a problematic access (ablation regions can be very rough), it is virtually impossible to built up and maintain such a tower. The alternative is a lower mast in combination with a cable balloon. Such a set-up was used in the meteorological experiment on the Pasterze, a 10 km long glacier in the Austrian Alps (PASTEX-94; Greuell et al. 1994). PASTEX-94 was carried out by research groups from the IMAU (Utrecht University) and the Department of Earth Sciences (Free University of Amsterdam) from mid-June to mid-August 1994. Six energy-balance stations were set up along the flow line of the glacier at elevations ranging from 2075 to 3225 m. Mass balance and different components of radiation were measured at all the stations. An analysis of the energy budget at the different stations can be found in Greuell et al. (1997).

To obtain information about the vertical structure of the glacier wind, at one station (A1, 2205 m above sea level) more extensive measurements were carried out. Here eddy-correlation measurements were made at two heights and a 13 m mast was erected with temperature, humidity and wind measurements at eight levels (0.25, 0.5, 1, 2, 4, 6, 8 and 13 m).

From this site soundings were made with a cable balloon up to a height of 500–1500 m above the glacier surface. The balloon measured temperature, humidity, windspeed and wind direction with a sampling time of 10 s. In total about 200 soundings were made.

Altogether, PASTEX-94 has delivered a unique data set on the structure of the katabatic flow over a melting valley glacier. We refer to Denby (1996), Van den Broeke (1997), Smeets et al. (1998) and Denby and Greuell (2000) for analysis of these data in various ways. Here we show a few subsets of the data that are typical for the glacier wind as generally observed on valley glaciers.

Figure 4 shows for a 3-d period half-hourly mean values of the windspeed at three heights on the profile mast. This is a period of dominating katabatic flow, which follows from the fact that the windspeed at 13 m is almost always lower than the windspeed at 6 m. The wind maximum is therefore below 13 m. It can also be seen that for very low windspeeds the flow at 1 m can be stronger than the flow at 13 m. The glacier wind is strongest in the afternoon, which is a consequence of the larger temperature deficit.

Data from two balloon soundings on Julian day 210 (27 July 1994) reveal the vertical structure of the circulation in the valley (Fig. 5). We first consider the sounding at 10:32 UT. All data samples (every 10 s) are shown. To resolve the large gradients in temperature and windspeed in the lower layers, the speed of ascent was kept low during the first few minutes of the sounding. Therefore the density of the measurements is higher in the lowest 50 m. The temperature inversion is about 12 °C. If we define the katabatic layer as the layer with a significant temperature deficit, its depth would be about 30 m. The lapse rate in the air above the katabatic layer is -0.0027 K m^{-1}. The highest wind speed is at the surface ($\sim$5 m s^{-1}), the wind direction is downglacier (the glacier flows in a southeasterly direction). It is clear that the katabatic layer is embedded in a valley wind of typically 2 m s^{-1}, flowing upglacier. At 16:27 UT the picture is rather similar. Temperature has increased in the lower few hundred meters, with little change higher up. Consequently, the lapse rate has now changed to about -0.0058 K m^{-1}. The balloon soundings do not resolve the structure of the katabatic flow in the lower 15 m. This, however, is obtained from the profile mast. Figure 6 shows wind and temperature profiles from the mast for the times corresponding to the beginning of the balloon soundings. The data points shown are half-hourly mean values.

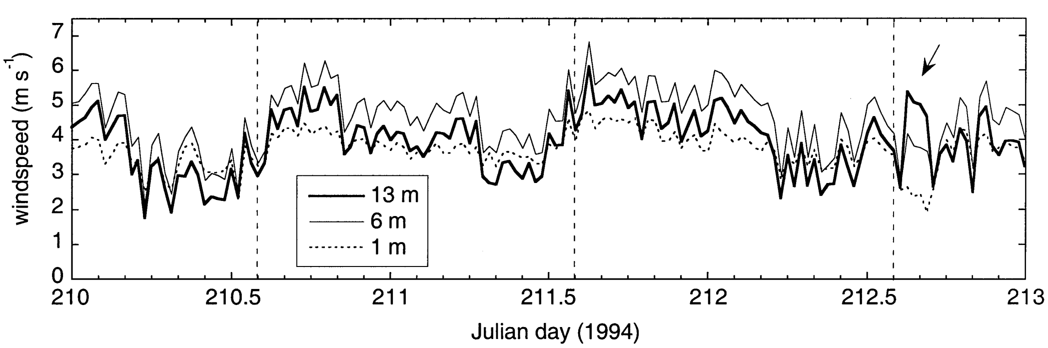

Fig. 4. Wind speed at some selected heights as measured on the tongue of the Pasterze glacier during PASTEX-94 (Greuell et al., 1994). The vertical dashed lines indicate local noon. The arrow points to a short period with the highest wind speed at 13 m.

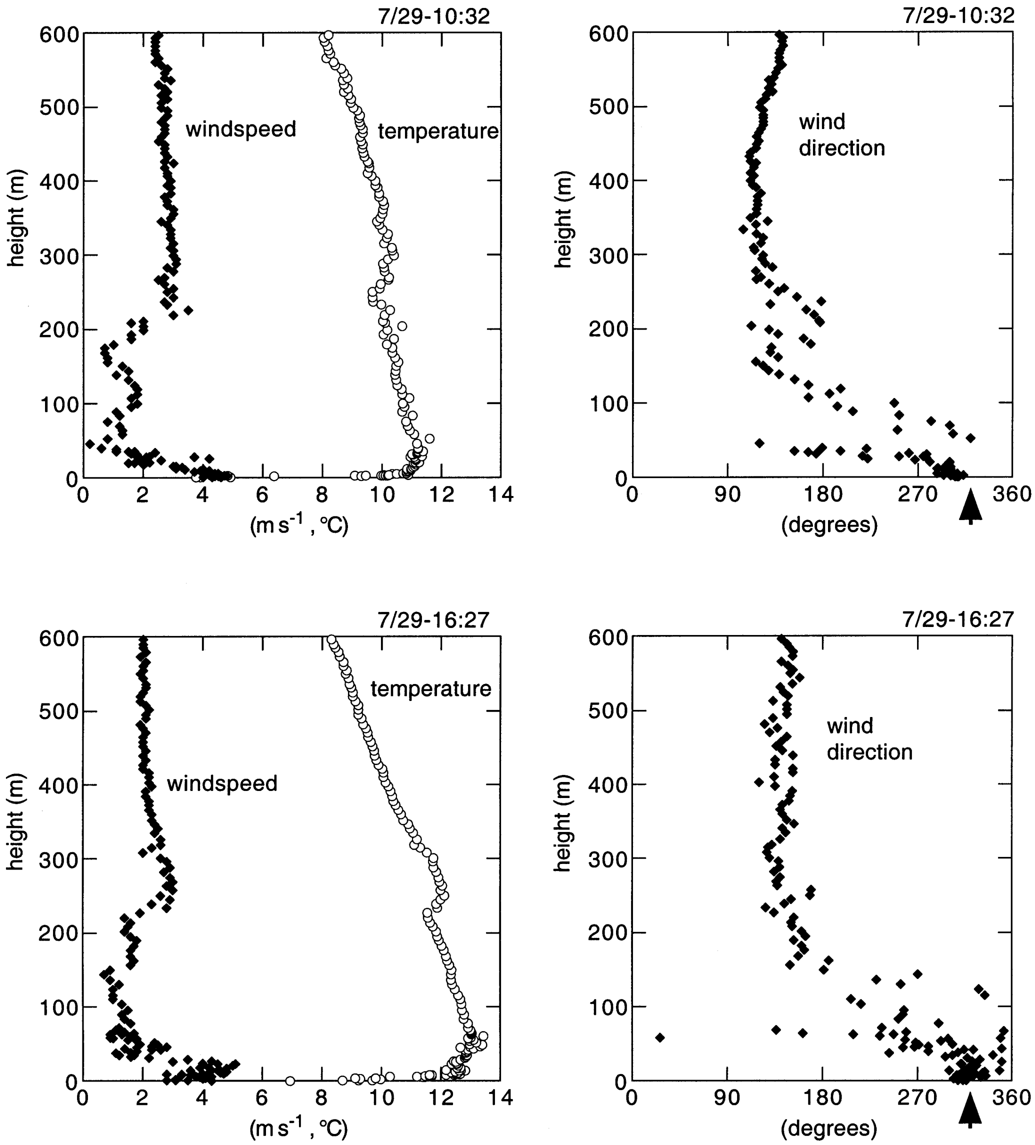

Fig. 5. Vertical structure of the glacier wind and valley circulation over the Pasterze glacier on the 29th of July 1994, measured with a cable balloon. The sample interval was 10 s. All samples are shown.

At 10:30 UT the wind maximum is found at a height of 2 m only. The maximum wind speed is 3.5 m s^{-1}, which is lower than the values suggested by the balloon sounding. We should keep in mind, however, that the balloon sounding does not yield half-hourly mean values. At 16:30 UT the wind maximum is at a height of 6 m and the wind speed here is slightly more than 5 m s^{-1}.

As has been reported already a long time ago (e.g. Hoinkes, 1954; Ohata and Higuchi, 1979), the height of the wind maximum in shallow katabatic flows tends to be larger when the flow is stronger. The extensive data set acquired during PASTEX-94 confirms this idea. Denby (1996) analysed data for a 2-wk period in which glacier wind occurred during 95% of the time. Because the profile mast had sensors at eight levels, the

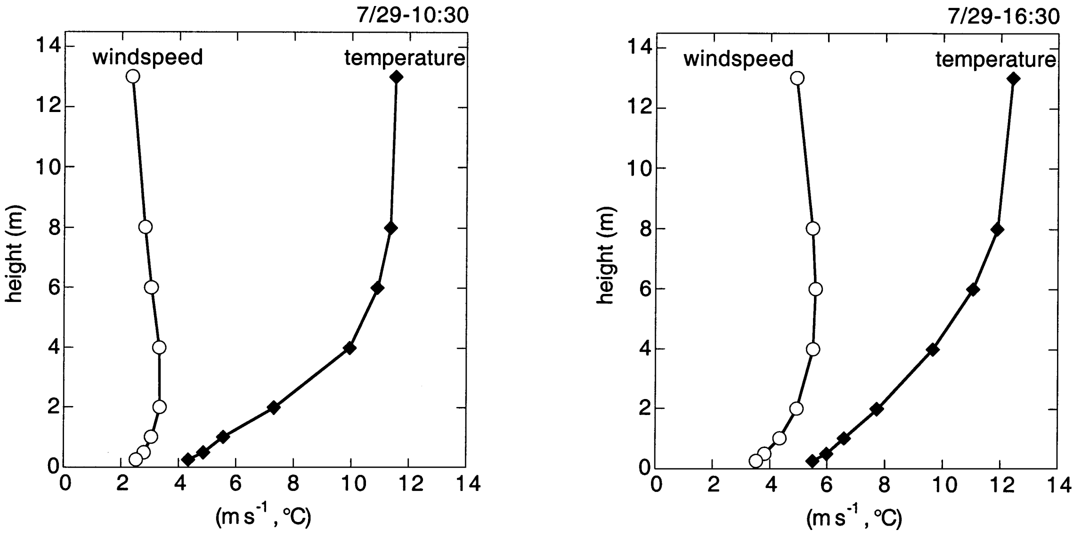

Fig. 6. Vertical structure of the glacier wind in the lower metres, measured with a profile mast. Each data point represents a 30-min average value. Date and time corresponds with the beginning of the balloon ascents shown in Fig. 5.

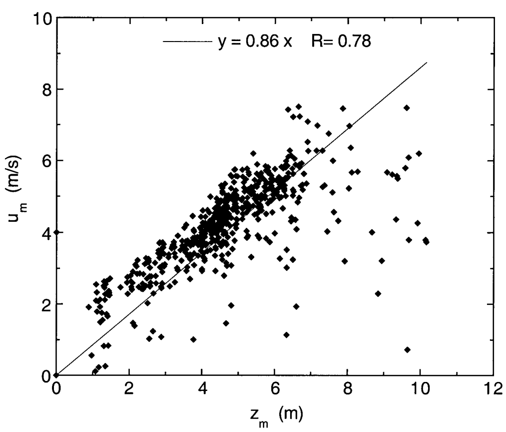

Fig. 7. Scatter plot of height of the wind maximum (z_m) against strength of the wind maximum (u_m) for the glacier wind on the Pasterze glacier (data prepared by Bruce Denby).

height of the wind maximum (z_m) can be determined accurately by polynomial fitting to the 30-min mean data values. The resulting relation between z_m and the maximum wind speed (u_m) is shown in Fig. 7. Any realistic model describing the vertical structure of the glacier wind should be able to reproduce the relation between u_m and z_m.

4. A simple theory to estimate the katabatic heat flux

In this section we consider the relation between the surface energy flux and the structure of the ambient atmosphere for cases with a glacier wind. In the classical scaling approach for the planetary boundary layer it is assumed that the driving term in the momentum equation, which is normally the large-scale horizontal pressure gradient, can be considered constant when the surface is approached. This then leads to the concept of the constant flux layer. In the case of a glacier wind the normal scaling procedure is principally wrong, because the largest driving term in the momentum equation, the (negative) buoyancy term, increases all the way to the surface. In fact, the vertical gradient in the buoyancy forcing is largest just above the surface. This implies that methods to estimate the surface fluxes in katabatic flow should, in one way or another, retain the coupling between the thermodynamical and mechanical part of the flow (Oerlemans, 1998). We note that this is fundamentally different from making stability corrections, determined by the thermal structure, in standard profile analysis.

The simplest self-consistent analytical model describing the glacier wind is the Prandtl model (Prandtl, 1942). Because this model is at the basis of our further discussion, we summarise it briefly here. It is based on the assumption that buoyancy forcing and friction are the leading terms in the downslope momentum budget. Moreover, it assumes that the sensible heat flux and vertical advection in the stably stratified atmosphere are the most important contributions to the heat budget. The analysis of Van den Broeke (1997) for the PASTEX-94 data certainly supports the hypothesis on which the Prandtl-model is based.

The governing equations are

$$\gamma u \sin(\alpha) - \frac{\mathrm{d}}{\mathrm{d}z}\left(K_h \frac{\mathrm{d}\theta}{\mathrm{d}z}\right) = 0 \quad \text{(heat balance)} \tag{1}$$

$$\frac{\theta}{T_0} g \sin(\alpha) + \frac{\mathrm{d}}{\mathrm{d}z}\left(K_m \frac{\mathrm{d}u}{\mathrm{d}z}\right) = 0 \quad \text{(momentum balance).} \tag{2}$$

These equations apply to a coordinate system in which the z-axis is perpendicular to the glacier surface, which has a slope α. The dependent variables are the downslope wind component (u) and the temperature deficit (θ). Furthermore, γ is the background potential temperature lapse rate, g the gravity acceleration, K_h the eddy diffusivity for heat, and K_m the eddy diffusivity for momentum. Prandtl (1942) solved these equations for constant eddy diffusivities and simple boundary conditions $[\theta(0) = C, \theta(z \to \infty) = 0, u(0) = 0, u(z \to \infty) = 0]$, in which C is the temperature deficit at the glacier surface. The solution reads

$$\theta(z) = C\mathrm{e}^{-z/\lambda} \cos(z/\lambda) \tag{3}$$

$$u(z) = -C\mu\mathrm{e}^{-z/\lambda} \sin(z/\lambda) \tag{4}$$

where

$$\lambda = \left(\frac{4T_0 K_m K_h}{\gamma g \sin^2(\alpha)}\right)^{1/4} \tag{5}$$

$$\mu = \left(\frac{g K_h}{T_0 \gamma K_m}\right)^{1/2}. \tag{6}$$

Here λ is the intrinsic length scale of the katabatic flow. The solution is shown in Fig. 8. Although there are a few serious problems with the Prandtl model to be discussed later, one must admit that the basic characteristics of the katabatic flow are captured well by the solution in eqs. (3) and (4).

The Prandtl model must be critisiced for a number of points. Because eddy diffusivity does not decrease when the surface is approached, the steep velocity and temperature gradients close to the surface do not appear in the solution. (Oerlemans, 1998) suggested to solving this problem by restricting the validity of the Prandtl model to the region above a surface layer. The depth of this surface layer is a fraction δ of the katabatic length scale λ. Oerlemans obtained good fits to the PASTEX-94 profile data with $\delta = 0.25$. However, to obtain surface fluxes it then has to be assumed that this surface layer is a constant flux layer. This may be a reasonable assumption for the heat flux, but probably not for the momentum flux (see also Fig. 8).

A more elegant approach is to let K_h and K_m vary with z, and use the WKB method to solve for the velocity and temperature profiles (Grisogono and Oerlemans, 2001a,b). By chosing appropriate profiles

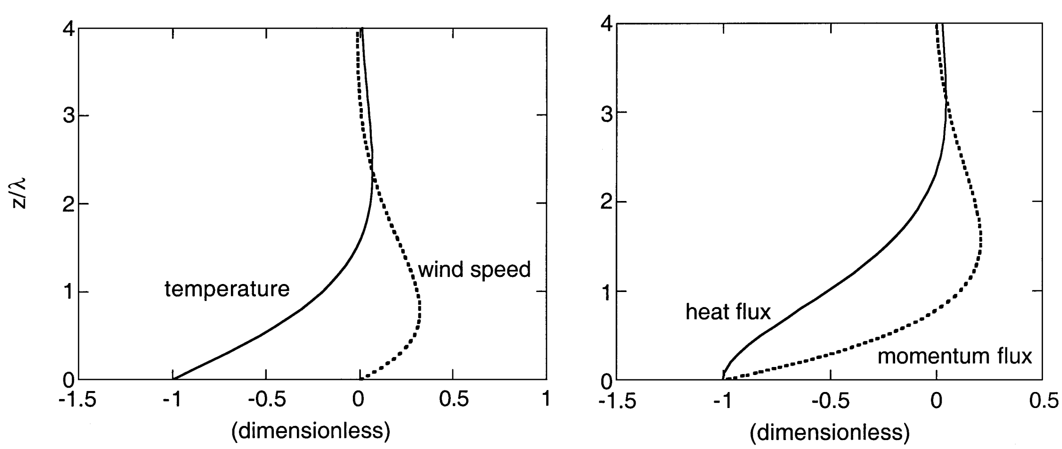

Fig. 8. Solution of the classical Prandtl model for katabatic flow.

for K_h and K_m, in which these quantities increase with z in the lower metres of the katabatic flow, a better fit to the observed wind and temperature profiles is obtained. Notably, the steep gradients near the surface are now better reproduced.

Another deficiency of the Prandtl model is the fact that the height of the wind maximum does not increase with the maximum windspeed (i.e. z_m is independent of u_m). Oerlemans (1998) proposed to improve the Prandtl model on this point by introducing flow-dependent eddy diffusivities, namely by making K_h and K_m proportional to λ (length scale) and u_m (velocity scale). This leads to a solution in which z_m and u_m are linearly related (in basic agreement with observations, see Fig. 7). Here we want to pursue this idea further, without referring explicitly to the Prandtl solution as given by eqs. (3) and (4).

The basic assumptions are that the heat and momentum balances are governed by eqs. (1) and (2), and that the flow is characterised by a well defined wind maximum. To characterise a 'katabatic state' we define scales for wind speed (u_s), temperature (θ_s) and length (z_s). We then obtain the following equations:

$$\gamma u_s \sin(\alpha) - \frac{K_h \theta_s}{z_s^2} = 0 \tag{7}$$

$$\frac{\theta_s}{T_0} g \sin(\alpha) + \frac{Pr K_h u_s}{z_s^2} = 0 \tag{8}$$

$$K_h = k z_s u_s. \tag{9}$$

The eddy Prandtl number (K_m/K_h) is denoted by Pr, and k is a dimensionless constant. Equations (7)–(9) are easily solved for u_s and z_s:

$$u_s = \theta_s \left(\frac{g}{T_0 \gamma Pr} \right)^{1/2}, \quad z_s = \frac{k\theta_s}{\gamma \sin(\alpha)}. \tag{10}$$

Next we chose the katabatic scales proportional to the maximum wind speed (u_m), the temperature deficit at the glacier surface (C), and the height of the wind maximum (z_m), respectively:

$$u_s = k_1 u_m; \quad \theta_s = -k_2 C; \quad z_s = k_3 z_m. \tag{11}$$

The constants k_1, k_2 and k_3 are of order unity. We then obtain from eq. (10)

$$u_m = -\frac{k_2}{k_1} C \left(\frac{g}{T_0 \gamma Pr} \right)^{1/2}, \quad z_m = -\frac{k_2 k}{k_3} \frac{C}{\gamma \sin(\alpha)}. \tag{12}$$

So according to this model u_m and z_m increase linearly with the forcing $-C$, which is in agreement with the observations (note that in the original Prandtl model the maximum wind speed increases with $-C$, while the height of the wind maximum is constant). The sensible heat flux at the glacier surface (katabatic velocity scale times katabatic temperature scale) becomes:

$$F_h = -k k_2^2 C^2 \left(\frac{g}{T_0 \gamma Pr} \right)^{1/2}. \tag{13}$$

The 'heat pump effect', as mentioned in the introduction, shows up clearly: the sensible heat flux increases quadratically with the surface temperature deficit C.

We have obtained simple expressions for the bulk properties of the glacier wind, but we are left with a large number of empirical constants (k, k_1, k_2, k_3). Currently available data do not allow a straightforward determination of these constants. If we want to obtain reasonable values for u_m, z_m and F_h, three empricial constants should be sufficient. Therefore we set $k_2 = 1$, implying that the katabatic temperature scale is just $-C$. Eddy correlation measurements of the turbulent fluxes during PASTEX-94 have shown that, for $C \approx -10$ K and $\gamma \approx 0.005$ K m^{-1}, F_h is typically 50 W m^{-2}. For $Pr \approx 5$, this yields $k \approx 0.0004$. Values of u_m and z_m in agreement with the PASTEX-94 data are obtained with $k_1 = 4$ and $k_3 = 2.5$. The results of this theory, with the values for (k, k_1, k_2, k_3) as given above, are summarized in Fig. 9.

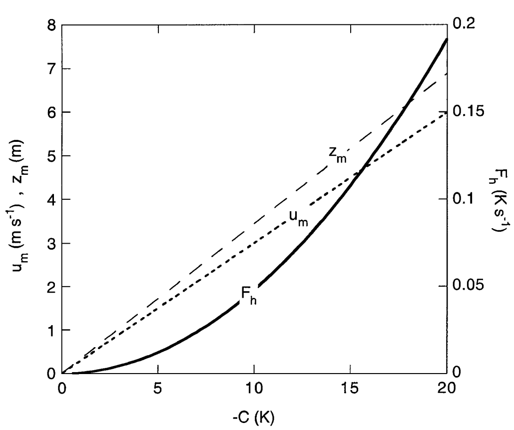

Fig. 9. Height and strength of the wind maximum and associated surface sensible heat flux in relation to the forcing (temperature deficit) C, as obtained from the model proposed in this paper.

Equation (13) can be used to parameterise the sensible heat flux in glacier mass-balance models. In fact, it can also be used to estimate the latent heat flux by defining a 'katabatic bulk exchange coefficient, K_{kat}' and assume that this exchange coefficient is the same for sensible and latent heat. To be consistent with eq. (13), K_{kat} should read:

$$K_{kat} = -kk_2^2 C \left(\frac{g}{T_0 \gamma Pr} \right)^{1/2} \quad (14)$$

The latent heat flux can then be estimated as

$$H_{la} = 0.622 \rho L_v K_{kat}(e_{air} - e_{surf}), \quad (15)$$

where ρ is the air density, L_v is the latent heat of vapourisation, e_{surf} is the vapour pressure at the surface, and e_{air} is the vapour pressure at a suitable level in the atmosphere. In spite of the limited possibilities to use AWS data for validation of the heat-flux parameterisation, we show one comparison that provides a bit of support. Oerlemans and Klok (personal communication) analysed AWS data from the Morteratschgletscher for the year 2000. They calculated the heat flux with the bulk method using surface temperature, air temperature (at 3.5 m) and wind speed (at 3.5 m) as input. The turbulent exchange coefficient for sensible and latent heat (constant for the whole year) was determined in such a way that the total ice melt calculated from the measured heat balance was exactly the same as the observed melt. A scatter plot of the 30-min values of the sensible heat flux (towards the surface) against air-surface temperature difference ΔT is shown in Fig. 10. Only the values with $\Delta T > 0$ are shown, but this comprises more than 95% of all data. Also shown is eq. (13) multiplied by ρc_p (set to 1000 J K^{-1}), with the parameter values mentioned earlier. C has been replaced by ΔT. There is a reasonable agreement, but the scatter is very large. A better fit can hardly be expected because ΔT is not more than a crude estimate of C. Also, the lapse rate of the ambient atmosphere is taken constant. The first-order agreement basically reflects the fact that in the observations wind speed generally increases with ΔT. Although Fig. 10 is encouraging, it is no proof of the validity of eq. (13). Our thrust is based on the fact that the simple theoretical model reproduces the basic characteristics of katabatic flow, notably the relation between the strength and the height of the wind maximum.

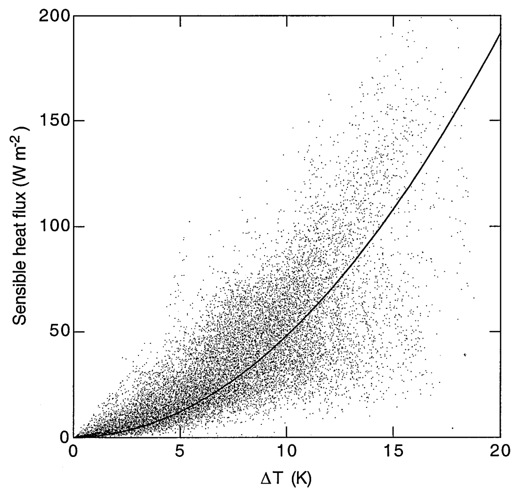

Fig. 10. A comparison between observed sensible heat flux (obtained with the bulk method) for the year 2000 on the Morteratschgletscher and the theoretical prediction [solid curve, eq. (13)].

5. Discussion

Data from automatic weather stations operated on glaciers have made clear that most of the time the flow is of a katabatic origin. Wind maxima occur at low levels and in spite of the very stable stratification, turbulence is generated and maintains a significant downward flux of sensible heat. The katabatic forcing increases towards the surface, which makes the applicability of standard MO theory doubtful.

Glacier mass-balance models contain a strongly parameterised scheme for the surface energy flux. Such models are normally forced by meteorological data from nearby weather stations, or by output from GCMs or numerical weather prediction models. In general, meteorological models do not have sufficient resolution to simulate the wind field over glaciers in mountanous terrain. For the Morteratschgletscher, for instance, a proper simulation of the wind field would require a horizontal resolution of typically 25 m and a vertical resolution of the order of 1 m (to resolve the glacier wind). Temperature, however, has a more robust mesoscale structure. Daily mean temperatures for the AWS on the Morteratschgletscher have been shown to correlate very well with temperatures measured at nearby weather stations, either in the valley or on a mountain top (Oerelemans, 2001).

The theory proposed in this paper can be used to estimate the sensible heat flux without explicit knowledge of the wind field, simply because the wind speed is obtained from a katabatic flow model. Clearly, our theory is most useful for small and medium-size glaciers. For larger glaciers, e.g. the Greenland ice sheet, atmospheric models will resolve the katabatic flow, and the effect of katabatic flow on the surface heat flux is therefore obtained directly (e.g. Denby et al., this volume).

For the practical case of modelling the surface energy flux over a small glacier, we propose the use of a bulk formulation that relates the surface heat flux to the ambient air temperature. The bulk exchange coefficient can be taken as the sum of K_{kat} and an 'external' component related to turbulence generated by the large-scale wind field.

The height of the wind maximum increases when the surface slope gets smaller [eq. (12)], which has been observed in many katabatic flows (including nocturnal drainage flows). However, K_{kat} is independent of the surface slope. In a more stable atmosphere (γ larger) K_{kat} is smaller. This is a dynamical effect (weaker katabatic flow) which should not be confused with the effect of stratification as modelled in the Monin–Obukhov theory.

Currently available data sets are not very suitable to test eq. (13). Strictly speaking, the AWS data do not provide an estimate of C because the height of the temperature sensor is too low (i.e. it is in the katabatic layer). Another problem is to make an estimate of the stratification of the ambient atmosphere (γ). In fact, this requires profiling with a balloon or with remote sensing techniques. Then reliable eddy correlation measurements of the heat flux are needed, but this is not a trivial matter in katabatic flows. Some of the data obtained during the summer experiments on Breidamerkurjökull and on the Pasterze (Smeets et al., 1998) may be used, but in fact a dedicated field experiment with turbulence measurements at different heights (also above the jet) would be needed for a rigorous test of the theory proposed in this paper.

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