The Use of Bulk and Profile Methods for Determining Surface Heat Fluxes in the Presence of Glacier Winds

A 1-D second-order closure model and in situ observations on a melting glacier surface are used to investigate the suitability of bulk and profile methods for determining turbulent fluxes in the presence of the katabatic wind speed maximum associated with glacier winds. The results show that profile methods severely underestimate turbulent fluxes when a wind speed maximum is present. The bulk method, on the other hand, only slightly overestimates the turbulent heat flux in the entire region below the wind speed maximum and is thus much more appropriate for use on sloping glacier surfaces where katabatic winds dominate and wind speed maximaums are just a few meters above the surface.
3.1 Introduction

The mass and energy balance of glaciers and ice caps are to varying degrees determined by the turbulent sensible and latent heat fluxes. Measurements made on glacier surfaces show that turbulent fluxes contribute from 20% to 40% of the total energy balance, see for example Kuhn (1979), Ohata (1989), van den Broeke (1996), Oerlemans et al. (1999) and Wagnon et al. (1999). They are thus an important component especially in relation to the climate sensitivity of the mass balance since it is these energy fluxes, along with longwave radiation, which are directly affected by temperature changes. It is therefore important in energy balance studies to correctly measure and model these fluxes.

Determining turbulent fluxes in the field can be done in several ways. The first is by direct eddy correlation methods made with sonic anemometers. Measurements using these instruments are the most direct, but also the most difficult to carry out chiefly due to the fragility of the instruments, the continuous maintenance and the problem involved with interpretation. The various attempts to directly measure turbulent fluxes on glacier surfaces, e.g. Munro (1989), Smeets et al. (1998) and van der Avoird and Duynkerke (1999), are generally only for short periods of time and do not always agree with other methods for determining the fluxes. The second, and most widely used method is to make use of mean wind, temperature and humidity measurements (made with robust instruments) and convert these mean values to surface fluxes by way of Monin-Obukhov (M-O) similarity theory. The third method, and most inaccurate, is the residual method whereby all the other components of the energy balance are measured, along with the mass balance, and the remaining melt must be explained by way of the turbulent heat fluxes. This is a useful check but errors in radiation measurements are generally so large that it cannot be used to determine these fluxes to any sufficient accuracy.

The second method described above, using mean values and M-O similarity theory, is the subject of this paper. Recent work with second-order models (Denby, 1999), which describe in more detail the turbulent structure of the atmospheric boundary layer (ABL) above sloping terrain has shown that assumptions made in M-O theory are often invalid on sloping glacier surfaces. This is the result of katabatic forcing in the ABL, which produces the katabatic, or glacier, wind. Katabatic winds are gravity flows caused by the turbulent cooling of air close to the surface. This cooler denser air sinks downslope producing the well known glacier wind, which is characterized by a low level wind speed maximum. The wind speed maximum found on glaciers with slopes of around 5° can be as low as 2 m above the surface and are thus close to or even below standard measuring heights. As is pointed out in Section 3.2, measurements made on the Pasterze glacier show wind maxima below 13 m for more than 75% of the time.

The presence of the wind speed maximum alters the turbulent scaling laws, used in M-O similarity theory, for two reasons. The first is the non-negligible turbulent transport term in the turbulent kinetic energy (TKE) budget. Since mechanical production of turbulence by shear is zero at the height of the wind speed maximum ($\partial U / \partial z = 0$) turbulent transport, a second-order term in the TKE budget, will dominate in this region and thus similarity arguments, which assume this term to be negligible, cannot be used. The second reason is that M-O similarity theory assumes
a constant turbulent flux layer, the so called surface layer where fluxes change in the vertical by less than 10%. This layer can be very thin when a katabatic wind speed maximum is present. As mentioned above, wind shear reduces to zero at the height of the wind speed maximum as will the vertical flux of horizontal momentum ($\overline{\text{uw}}$). This indicates a strong divergence in $\overline{\text{uw}}$ and thus a very thin surface-layer. For a wind speed maximum of 5 m M-O theory would then only be valid in the lowest 0.5 m.

In spite of this, M-O theory is still used to calculate turbulent fluxes under these conditions. The question posed here is what sort of influence the presence of a wind speed maximum has on the bulk and profile methods commonly used to determine these fluxes. By making use of observations and second-order modelling results it will be shown that profile methods will always severely underestimate turbulent fluxes when a wind speed maximum is present but that the bulk method still gives quite good estimates of these fluxes up to at least the height of the wind speed maximum.

### 3.2 Observations

In this paper use is made of experimental data from the PASTEX glacia-meteorological experiment carried out on the Pasterze glacier in Austria during the summer of 1994 (Greuell et al., 1997). Six weather stations for determining the energy balance of the glacier were stationed along the central flow line. Most of these consisted of just two measurement heights, 0.5 m and 2 m, but at one site, known as A1, a 13 m profile mast with 8 measurement heights (0.4, 0.7, 1, 2, 4, 6, 8 and 13 m) and a balloon sounder were placed at a point roughly 1 km from the end of the glacier tongue at an elevation of 2200 m a.s.l.. Smeets et al. (1998) provides more information concerning the measurements made using this mast. The local slope at this site was approximately 3.5°.

During the observational period from 16 June to 11 August over 2600 half-hourly average observations were made. Of these 77% showed the existence of a wind speed maximum below the highest profile level when wind direction was downslope, indicative of katabatic flows. Air temperatures were well above 0 °C during the observational period and the surface was almost always melting. Figure 3.1b shows the average downslope wind speed and temperature profiles during a two day fair weather period at this site. During this period a wind speed maximum was present more than 90% of the time.

### 3.3 The mean and turbulent structure of katabatic flows

Katabatic flows on glaciers are driven from beneath by buoyancy, or katabatic, forcing. Warm air overlaying the glacier is cooled by turbulent exchange with the ice or snow surface, becoming denser and resulting in the downslope flow of cool air. This katabatic flow, often referred to as glacier wind, is a dominant feature of the ABL above most temperate glaciers. The low level forcing combined with the turbulent exchange of momentum will intrinsically lead to the development of a wind speed
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A 1-D second-order turbulence closure model, which simulates the vertical profile of the ABL, is used here to help describe the turbulent structure of katabatic flows. The 1-D model is discussed in detail in Denby (1999) where it has been shown to describe the mean and turbulent structure of katabatic flows quite well. It makes use of second-order closures, including closures for the turbulent transport of TKE, thus M-O similarity is not assumed in order to calculate the katabatic profiles. Only at the lowest level of the model, \( z = 0.2 \text{ m} \), is the assumption of similarity used to calculate lowest level fluxes. The roughness length for momentum in the model is predefined and the roughness length for temperature is based on the surface renewal model from Andreas (1987). The flux profile relationships, as defined in Section 3.4, are used for the stability correction at this lowest level.

Equations 3.1 and 3.2 show the simplified 1-D equation for downslope momentum \( U \) and temperature perturbation \( \Theta \).

\[
\frac{\partial U}{\partial t} = -\frac{\partial \overline{w'w'}}{\partial z} + \sin(\alpha)\frac{\partial}{\partial \Theta} + \cos(\alpha) f(V - V_g) \quad (3.1)
\]

\[
\frac{\partial \Theta}{\partial t} = \sin(\alpha) \gamma \theta U - \frac{\partial \overline{w'\theta'}}{\partial z} \quad (3.2)
\]
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Figure 3.2: a) Simulated downslope wind speed (U) and temperature deficit (Θ) profiles for the 1-D katabatic case described in the text. The wind speed maximum is at a height of 5 m. b) Simulated uw and wθ profiles. c) Simulated TKE budget profile.

Θ is defined as the potential temperature difference from a pre-defined ambient temperature profile given as \( \Theta(x_3) = \Theta_o + \gamma_o x_3 \), where \( x_3 \) is the true vertical component, i.e. parallel with the gravitational vector, and where the co-ordinate system \((x, z)\) is defined as a cartesian co-ordinate system orthogonal to the sloping surface with slope angle \( \alpha \). \( V_g \) is the geostrophic wind component and the over bar indicates the turbulent flux terms, as defined by Reynolds decomposition. These are the vertical flux of horizontal momentum \( \bar{uw} \) and the vertical flux of potential temperature \( \bar{w\theta} \), which is directly converted to the turbulent flux of sensible heat \( H = \rho c_p \bar{w\theta} \) where \( \rho \) is the density and \( c_p \) the specific heat of air. A schematic representation of the glacier wind is given in Figure 3.1a indicating the terms mentioned above.

In Figure 3.2a the simulated downslope wind and temperature profiles are shown for a typical case of katabatic forcing on a sloping glacier with surface temperature perturbation \( \Theta_s = -10 \, K \), surface roughness length \( z_o = 2 \, mm \), temperature lapse rate \( \gamma_o = 3 \, K \, km^{-1} \) and slope \( \sin\alpha = -0.1 \, (6') \). The major features of the mean flow are of course the wind speed maximum, in this case at a height of around 5 m, and the strong temperature inversion beneath the wind speed maximum of approximately \( 0.5 \, K \, m^{-1} \).

On typical temperate glaciers katabatic forcing is the dominant driving force in the momentum budget close to the surface (Denby and Smeets, 2000). This is balanced by the flux divergence of horizontal momentum in Equation 3.1 resulting in the strong gradient in \( \bar{uw} \) seen in Figure 3.2b. \( \bar{uw} \) will pass through zero at, or at least close to, the height of the wind maximum. Though there is a steep gradient in \( \bar{uw} \), the vertical profile for sensible heat flux \( \bar{w\theta} \) is almost constant in the region below the wind speed maximum, varying in this case by just 15% (Figure 3.2b). Though this is a 1-D simulation and other terms in the temperature budget are also important in a 2- or 3-dimensional case, such as horizontal advection of temperature perturbation, the basic turbulence structure remains the same.

In order to indicate the role played by turbulent transport in katabatic flow the TKE budget is also shown in Figure 3.2c. As the height of the wind speed maximum
is approached the mechanical production through shear approaches zero and the turbulent transport term transports TKE into this region. This is where M-O theory, as well as local scaling arguments, break down since they do not take into account the transport terms. Above the wind speed maximum, where shear dominates the TKE budget, local scaling arguments such as those described by Nieuwstadt (1984) become applicable again.

### 3.4 Suitability of bulk and profile methods

Before comparing the bulk and profile methods with observations and model results it is useful to give a brief description of these two methods. Both the bulk and profile methods use Monin-Obukhov similarity theory, which is based on scaling arguments using the turbulent velocity and temperature scales, $u_*$ and $\theta_*$, and the length scales $z$ (height), $L$ (Obukhov length), and $z_0$ and $z_h$ (surface roughness lengths of momentum and temperature). These turbulent scales define completely, under idealized horizontally homogeneous and quasi steady-state conditions, the turbulent structure of the surface layer, the lowest region of the atmospheric boundary layer (ABL).

The flux-profile relationships, Equations 3.3 and 3.4, are definitions relating the vertical gradients of mean wind and temperature to these turbulent scales:

$$
\phi_m = \frac{\kappa z}{u_*} \frac{\partial U}{\partial z} \quad (3.3)
$$

$$
\phi_h = \frac{\kappa z}{\theta_*} \frac{\partial \Theta}{\partial z} \quad (3.4)
$$

where $u_* = \sqrt{\tau \rho / \rho_0}$ and $\theta_* = \frac{\sqrt{\tau \rho_0}}{u_*}$.

Under neutral conditions $\phi_m$ is defined as being equal to unity and the von Kármán constant ($\kappa=0.4$) is determined from observations to fulfill this definition. Integration of these relationships, under stable conditions and the assumption of a constant flux layer, leads to the well known log-linear wind and temperature profiles if a linear relationship is assumed for $\phi_{m,h}$ of the following form (Garratt, 1992):

$$
\phi_m = 1 + \alpha_m \frac{z}{L} \quad (3.5)
$$

$$
\phi_h = Pr + \alpha_h \frac{z}{L} \quad (3.6)
$$

where $L = \frac{u^2}{\kappa (\tau / \rho_0)^{1/2}}$.

The constants in Equation 3.5 and 3.6 are determined solely from observations and vary to some extent in the literature. We use here a value of 5 for both $\alpha_m$ and $\alpha_h$ and a near-neutral turbulent Prandtl number $Pr=1$. Integrating Equations 3.3 and 3.4 from level $z_1$ to $z_2$ leads to the following log-linear profile equations:

$$
(U(z_2) - U(z_1)) \frac{K}{u_*} = \ln \left( \frac{z_2}{z_1} \right) + \alpha_m \frac{z_2 - z_1}{L} \quad (3.7)
$$
and integrating these equations from the roughness length heights of $z_o$ and $z_h$ to $z$ gives the log-linear bulk equations:

\[
\frac{U(z)}{z} = \ln \left( \frac{z}{z_o} \right) + \alpha_m \frac{z}{L} \quad (3.9)
\]

\[
\left( \Theta(z) - \Theta_s \right) \frac{K}{\theta_s} = Pr \ln \left( \frac{z}{z_h} \right) + \alpha_h \frac{z}{L} \quad (3.10)
\]

It should be noted that the roughness lengths are the height at which the extrapolated wind and temperature profiles would reach their surface values.

Equations 3.7 and 3.8 require a minimum of two profile heights for wind and temperature to derive the four unknowns $u_*, \theta_*, z_o$ and $z_h$ assuming the surface temperature $\theta_s$ to be known. For simplicity in this paper we will limit ourselves to the case where only two profile heights are available, this defines the profile method for determining surface fluxes. The bulk method on the other hand makes use of only one height to derive $u_*$ and $\theta_*$ assuming the roughness lengths ($z_o, z_h$) and the surface temperature $\theta_s$ to be known. The application of these methods has been quite successful in the stable ABL, which is the reason for its popularity under most conditions.

We will now take bulk and profile derived fluxes calculated from the profile mast data and compare these to model simulations. The results are presented in Figures 3.3 and 3.4 for both $\overline{uw}$ and $\overline{w\theta}$ where we have used the non-dimensional height $z/H$, $H$ is the height of the wind speed maximum, as the scaled vertical axis. In these figures all mast levels are used to calculate the bulk and profile derived fluxes. For the profile method each point represents the profile derived flux taken from two adjoining mast levels at a height determined by their average. For the bulk method each point indicates the bulk derived fluxes for a particular mast level. Both sets of data are normalized by the bulk derived value taken from the 0.7 m mast level using previously derived roughness lengths (Denby and Smeets, 2000). This is considered to give the best estimate of the surface flux since it is expected that when $H \gg z \gg z_o$, M-O theory will still be valid . The normalized fluxes should approach unity in this range.

The model results are calculated by a continuous run of the 1-D model where the surface temperature perturbation is allowed to slowly decrease by $-0.1$ K hr$^{-1}$ from 0 to $-20$ K. In this way a continuous range of simulated katabatic profiles are generated by varying the katabatic forcing. Fixed heights of 2 and 0.5 m within the model are then used to calculate the profile derived fluxes and a height of 2 m is used to calculate the bulk derived fluxes. Both the bulk and profile derived model fluxes are normalized by the model generated surface flux value. The local slope, $\alpha = 3.5^\circ$, is used as well as the average lapse rate of $\gamma_\theta = 3.5$ K km$^{-1}$.

It is worth noting at this point that even though the simulations are one dimensional the basic turbulent structure of the katabatic flow remains the same, even with the introduction of advection terms in Equations 3.1 and 3.2. As previously mentioned, Section 3.3, the momentum budget is dominated by the katabatic forcing.
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The temperature budget, on the other hand, is more strongly affected by the horizontal and vertical advection of temperature perturbation. To test the sensitivity of the results to temperature advection an entrainment velocity is introduced to the temperature budget, as was carried out in Denby (1999), which allows the representation of vertical advection within the 1-D model. Even under conditions of strong entrainment the normalized profiles shown in Figures 3.3 and 3.4 remain essentially

Figure 3.3: Profile derived normalized vertical momentum flux (a) and sensible heat flux (b) as function of the non-dimensional height \( z/H \) (\( H=\)height of the wind speed maximum) derived from observations, dots, and from 1-D model simulations, continuous line. Both the heat and momentum fluxes are normalized by the bulk derived surface flux determined at a height of 0.7 m. See text for details.

Figure 3.4: Bulk derived normalized vertical momentum flux (a) and sensible heat flux (b) as function of the non-dimensional height \( z/H \) (\( H=\)height of the wind speed maximum) derived from observations, dots, and from 1-D model simulations, continuous line. Both the heat and momentum fluxes are normalized by the bulk derived surface flux determined at a height of 0.7 m. See text for details.
Figure 3.5: Gradient Richardson number (a) and Bulk Richardson number (b) as function of the non-dimensional height $z/H$ ($H$=height of the wind speed maximum) derived from observations, dots, and from 1-D model simulations, continuous line. The dotted line in (a) indicates the model’s critical gradient Richardson number of 0.23. The bulk Richardson number is defined as $Ri_B = \frac{\bar{w}U_0 \bar{\theta}}{\bar{U}^2 + \bar{V}^2}$ unaltered. In this regard the 1-D results obtained here are quite robust.

The profile derived fluxes show a large amount of scatter in the observations, which results from the sensitivity of this method to measurement errors and variability. What is clear though is that turbulent fluxes are severely underestimated throughout the region below the wind speed maximum. This is the result of reduced shear and increasing temperature gradients as the wind speed maximum is approached. This ensures that the local gradient Richardson number ($Ri = \frac{\partial \bar{U} \partial \bar{\theta}}{(\partial \bar{U}/\partial z)^2 + (\partial \bar{V}/\partial z)^2}$) approaches its critical value ($Ri_c = 0.23$ in this model), above which all turbulence is suppressed according to M-O theory. This occurs at around half the height of the wind speed maximum (Figure 3.5a).

The bulk method, on the other hand, shows far less scatter since gradients need not be determined. Though $\overline{u \bar{\theta}}$ is underestimated when measurements are made above the height of the wind maximum this is not as severe as in the profile case. $\overline{w \bar{\theta}}$, on the other hand is slightly overestimated when measurements are made below the wind speed maximum but reduce quickly as the measurement height increases above the wind speed maximum. Generally wind maxima are above a normal measuring height of 2 m so we are not usually concerned with this region.

How can we explain the ability of the the bulk method to estimate the surface fluxes even in the region of the wind speed maximum? The bulk method is essentially an integrated form of the profile method and, as such, is less sensitive to variations in gradients brought about by the presence of a wind speed maximum. As turbulence decreases with decreasing shear near the wind maximum, wind speeds reduce relative to their log-linear form (Equation 3.9). This effect is shown in Figure 3.6, where the katabatic wind and temperature profiles from Figure 3.2 are plotted against logarithmic and log-linear profiles with equivalent surface flux values. This then leads to a reduction in the bulk derived $\overline{u \bar{\theta}}$ values in Figure 3.4a. How-
ever, this same reduction in turbulence will increase the temperature gradient, and therefore the temperature, as the wind speed maximum is approached (Figure 3.6b). These two effects tend to cancel each other out, reducing $u_*$ and increasing $\theta_*$, and the resulting bulk determined heat flux remains fairly constant until the diminishing wind speed above the wind maximum leads to stability corrections severely reducing the apparent turbulence.

For the 1-D katabatic case the normalized model profiles generated for Figure 3.6 are not universal. The height of the katabatic wind speed maximum ($H$) is not the only scaling length that determines the form of the normalized curves since the turbulent M-O scaling length $L$ is of a similar order. The dependence on slope for the bulk derived heat flux is illustrated in Figure 3.7a where three model runs with differing slopes are shown. From a pragmatic point of view the slope dependence is of little consequence since the largest deviations from unity occur by weaker slopes where wind maxima are higher and so measurements will always be made well below the wind speed maximum. It is therefore relieving to see that even for steep slopes, where the wind speed maximum can be quite low, that the bulk derived heat flux appears to be quite representative of the surface value.

To illustrate this point further, the height of the wind speed maximum is plotted as a function of maximum wind speed for the three differing slope angles in Figure 3.7b. As in all the simulations the surface temperature deficit is allowed to slowly decrease from 0 to -20 K and the closed circles in Figure 3.7b indicate 2 K steps. For wind speed maxima under 2 m the temperature perturbation and wind speed are quite low, corresponding to small surface turbulent heat fluxes, especially for lesser slopes. Any errors introduced due to the proximity of the wind speed maximum will be, on an absolute scale, negligible.
3.5 Conclusion and discussion

The suitability of Monin-Obukhov similarity theory under conditions of katabatic flow where a low level wind speed maximum is present is discussed. Though assumptions made in M-O theory are not valid in the presence of a wind speed maximum it is shown that bulk estimates of turbulent heat fluxes give quite reasonable results in the entire region below the wind maximum whereas profile derived fluxes underestimate the surface fluxes severely. Only in the lowest region, $z/H < 0.5$, does the profile method give any value for the surface flux at all, due to stability corrections, and only in a narrower region, $z/H < 0.3$, do we find values comparable to low level bulk estimates. For a wind speed maximum at 5 m this would infer that profile measurements would need to be made below a height of approximately 1.5 m. However, profile fits need to be made well above the average surface roughness element size which can easily be of the same order on some melting ice surfaces.

That bulk methods are effective in determining turbulent fluxes, even under conditions of katabatic flow, is good news for energy balance studies since measurements of mean wind and temperature at one level are far easier to carry out than profile or eddy correlation measurements. There remains, however, two problems in using the bulk method to determine turbulent heat fluxes. The first is the determination of the surface temperature. Though this can be an important point for non-melting surfaces it is inconsequential when studying melting ice surfaces, as is done here, since the surface is permanently at the melting point.

The second problem is the determination of surface roughness lengths for momentum, temperature and water vapour, which are essential parameters when using the bulk method. Normally roughness lengths are determined using profile methods since this is how they are defined, being the height at which the extrapolated