MODELING THE TURBULENT STRUCTURE OF THE KATABATIC JET

Stefan Söderberg* and Michael Tjernström
Stockholm University, Stockholm, Sweden

1. INTRODUCTION

Katabatic flows, also known as drainage or gravity flows, are formed when the air adjacent to a sloping surface cools relative to the air at the same absolute elevation away from the surface. The low-level downslope flow is triggered by negative buoyancy, which then acts also horizontally, the so-called katabatic forcing term. In general, this type of flow can form over any sloping surface. However, the katabatic forcing is often smaller than other terms in the momentum budget, such as the synoptic pressure gradient. Thus, katabatic flows are typically only observed during clear sky conditions, when nighttime radiative cooling of the surface peaks. Over icecaps on the other hand, the katabatic forcing term is in most cases appears strong enough to overcome the background pressure gradient term. Persistent katabatic flows over glaciers have been reported in numerous studies (e.g., Gruell et al. 1994; Oerlemans et al. 1999).

Katabatic flows drive the turbulent exchange of heat and momentum between the surface and the free atmosphere. Since glacier melting is most sensitive to changes is in long-wave radiation and the turbulent heat flux (e.g., Oerlemans, 2001), the properties of katabatic flows are important for the understanding of glacier response to climate change. In the present study we take a closer look at the modeled turbulence structure in a katabatic wind-speed jet over a melting glacier.

2. MODEL DESCRIPTION AND SETUP

The numerical model used in the study is the Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS™) version 2.0 atmospheric mesoscale model (Hodur, 1997). It is a nonhydrostatic compressible model with a terrain-following sigma-z vertical coordinate. Among the physical parameterizations in the model is a level 2.5 turbulence closure (Mellor and Yamada, 1982); Turbulent kinetic energy (TKE) is a prognostic variable while other second order moments are obtained from steady-state analytical expressions.

Nested grids were used with a 27-km horizontal resolution in a coarse mesh and a 3-km resolution in the innermost grid (Figure 1). To be able to resolve the shallow boundary layer flow studied here, 40 vertical model levels were used, with the lowest level at 1 m and 2-m intervals up to 15 m. Initial and lateral boundary conditions were provided using ECMWF analyses.

* ***Corresponding author address:*** **Stefan Söderberg, Deparment of Meteorology, Stockholm University, SE-106 91 Stockholm, Sweden; email: stefan@misu.su.se**

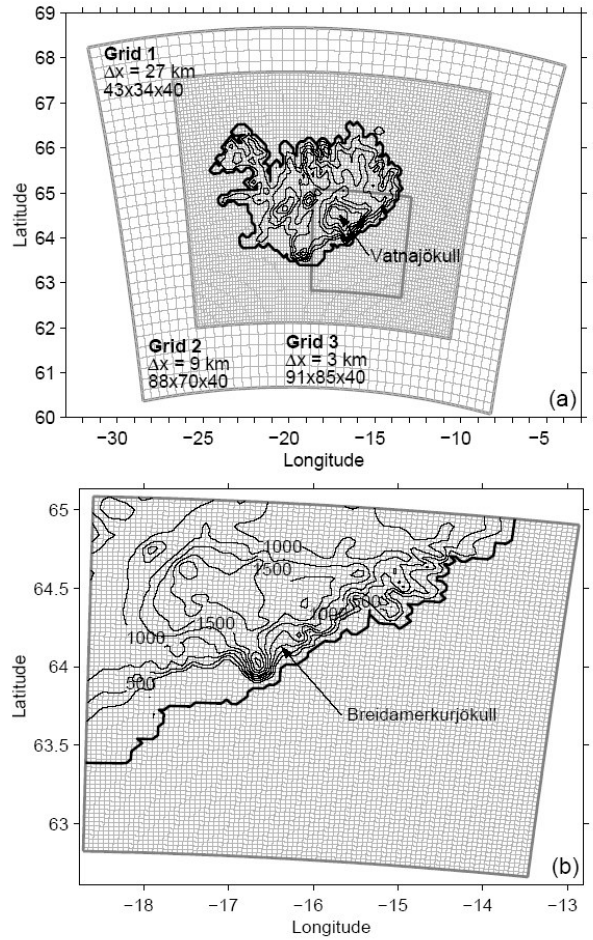


Figure 1. Horizontal model domains. Terrain elevation is contoured every 250 m, black bold line is coastline: a) coarse mesh (grid 1) and the two inner nests (grid 2 and 3); b) Innermost nest (grid 3).

3. RESULTS

Before studying the modeled turbulence structure of the katabatic jet, a brief description of the simulated mean boundary-layer structure over Breidamerkurjökull is given. Figure 2 shows the near-surface flow (~ 11 m above ground level) for a case from 7 July 1997, illustrated by wind vectors. Dashed contours show scalar wind speed and the solid contours show terrain elevation. From the top of the glacier and down towards the coastline, the flow accelerates to more than 4 m s^{-1} and turns into the glacier fall line. Note that in the upper part of the glacier and some distance offshore, the wind direction has a southerly component. A principal agreement between observations and model results is found (not shown).

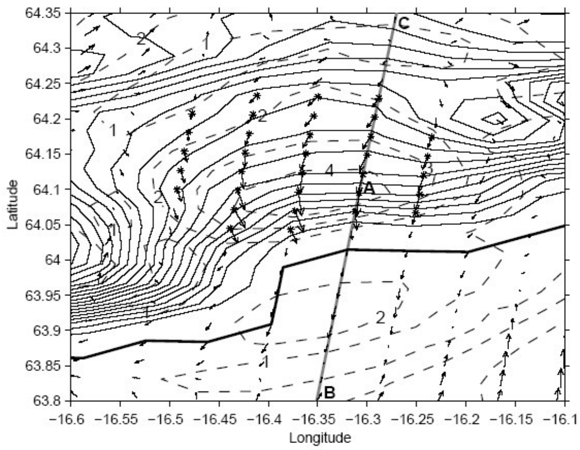


Figure 2. Scalar wind speed (dashed) and wind vectors 11 m agl on 7 July 12 UTC in the immediate surroundings of Breidamerkurjökull. Terrain elevation is contoured every 100 m; stars show grid points in which turbulence quantities are analyzed.

The spatial variation of the flow over Breidamerkurjöll is further illustrated in Figure 3, showing a vertical cross-section of temperature and the meridional wind-speed component taken along B-C in Figure 2. Although this cross-section is not along a trajectory, it provides a useful illustration of what an air parcel may experience on its way from the top of Breidamerkurjökull down towards the ocean. As the air parcel accelerates down the slope it undergoes an adiabatic heating. This means that the

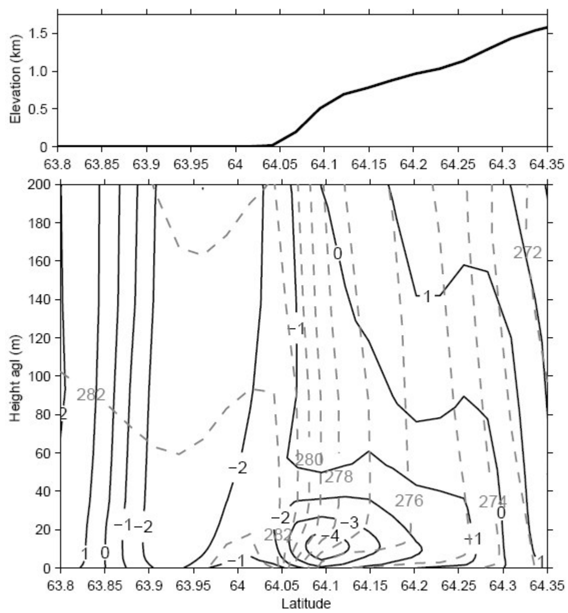


Figure 3. Cross section along B-C of Figure 2 of the meridional wind component (m s^{-1}, solid) and temperature (K, gray dashed). Vertical axis is in m above ground level. Terrain elevation in km above sea level along the cross-section is also shown. Negative values of the meridioanal wind component are here directed close to the glacier fall line.

strength of the surface temperature inversion will continuously increase, since the surface temperature remains constant at the melting point of the ice. In a sense, katabatic flows over a melting glacier can this way be considered self-sustaining, since the negative buoyancy is continuously reinforced by the flow itself. This could partly explain the observed persistence of katabatic flows over melting glaciers.

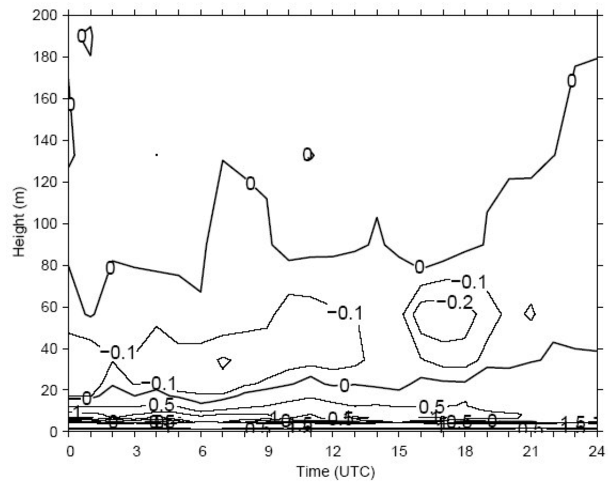


Figure 4. Contour plot of the Scorer parameter, L^2x10^{-2}, from the model grid point marked by A in Figure 2. Bold lines are the zero isolines.

From observations over Breidamerkurjökull, Parmhed et al. (2004) hypothesized that wave trapping could also be important for the persistence of katabatic flows. This was based on calculations of the Scorer parameter, L^2, which within linear theory is defined as (Nappo 2002):

$$L^2 = \frac{N^2}{U^2} - \frac{1}{U}\frac{\partial^2 U}{\partial z^2}\ , \qquad (1)$$

where U is flow speed and N the buoyancy frequency. Above the wind-speed jet, Parmhed et al. (2004) found negative values of L^2 implying that vertically propagating waves are not possible and instead waves decay exponentially with height. Small or negative values of L^2 are due to the wind-speed curvature term that becomes large compared to the buoyancy term, a characteristics of the jet itself. In Figure 4, L^2 calculated from model results is shown. This contour plot is typical for the lower part of Breidamerkurjökull. The deep layer with negative values of L^2 suggests persistent wave trapping above the katabatic jet and a decoupling of the near-surface flow from the ambient flow aloft. Wave trapping separating a marine boundary layer from the free atmosphere has also been discussed in conjunction with coastal wind-speed jets (Burk et al. 1999).

Profiles of modeled momentum flux over Breidamerkurjökull are shown in Figure 5. The height is here normalized by the jet height, z_{jet}, and the momentum flux is normalized by the square of the surface friction velocity. A striking feature is the high magnitude of upward directed momentum flux above the wind-speed jet. This is a

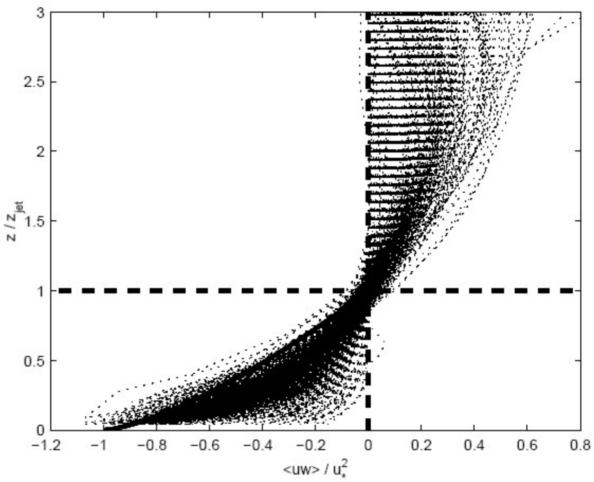


Figure 5. Modeled momentum flux normalized by the square of the surface friction velocity plotted against normalized altitude. Data points are from model grid points marked with stars in Figure 2, 00 to 24 UTC 7 July 1996. Solid line is the profile for stable stratification suggested by Lenshow et al. (1988).

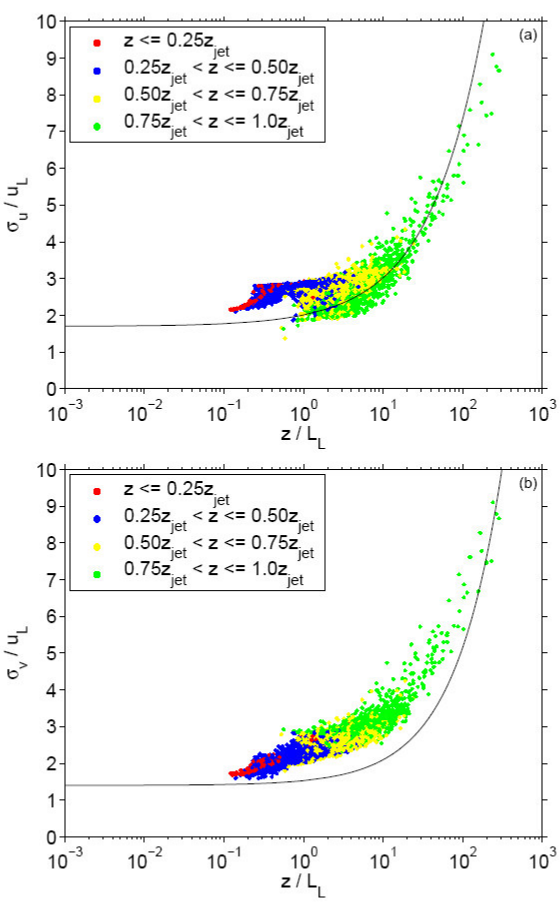


Figure 7. Modeled turbulence quantities over Breidamerkurjökull 00 UTC to 24 UTC 7 July 1996: (a) along-flow component, (b) cross-flow component. The scaled variances are partitioned into groups by normalized height according to legends.

distinct difference from stable boundary layers in general, but was also found in a coastal jet by Brooks et al. (2003). Below the wind speed maximum, the momentum flux profiles are more concave than the analytical expression (thick solid line) suggested by Lenshow et al. (1988) for stable stratification.

In the wind-speed jet, the gradient Richardson number has a distinct super-critical maximum (not shown). Nevertheless, the flow remains turbulent through the jet although shear production has by definition a distinct minimum at the wind-speed maximum (Figure 6). The turbulent transport terms provide a net flux of turbulence into the jet that keeps TKE from going to zero locally.

Local similarity scaling, with scales defined as (Niewstadt 1984):

$$u_L = \left(\overline{w'u'}^2 + \overline{w'v'}^2\right)^{0.25} \quad (2)$$

$$L_L = -u_L^3 \theta_v \big/ \left(\kappa g \overline{w'\theta'_v}\right), \quad (3)$$

was applied to the continuously turbulent boundary layer and the along-flow and cross-flow components are shown in Figure 7, along with the empirical functions found in a coastal environment (Brooks et al. 2003). Distinct features are the increased normalized velocity variances with increasing stability. This is presumably due to the non-locality induced by the fluxes of turbulence into the jet. The velocity components appear to follow a functional relationship similar to that from the coastal environment.

In contrast to measurements on other glaciers (e.g., Smeets et al. 1999), the scaled cross-flow component increases significantly at the wind speed maximum (Figure 8). We believe this to be a direct consequence of an apparent wind-direction shear across the jet. This tends to keep the shear-production from going to identically zero as there will be a non-zero wind-speed

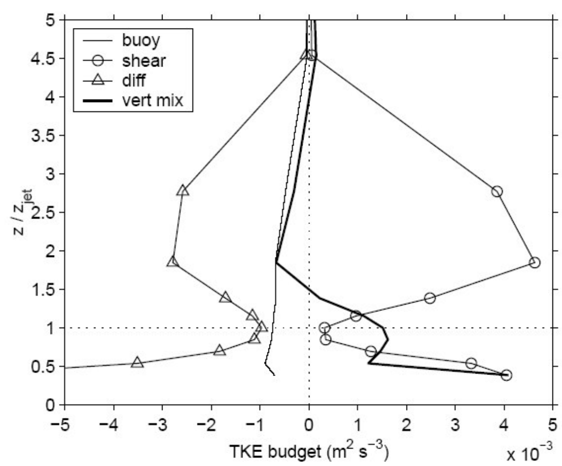
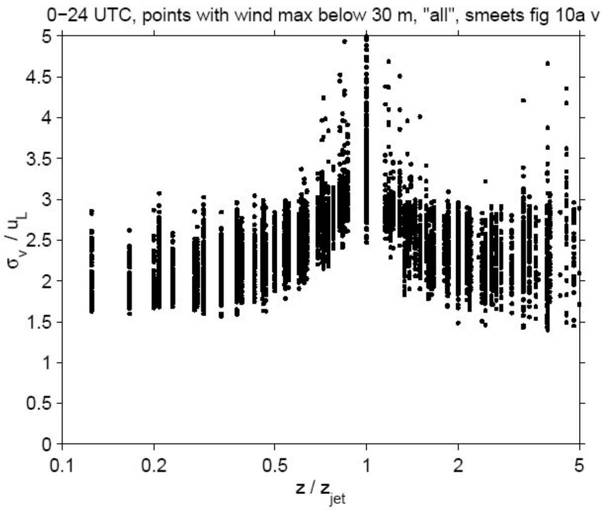


Figure 6. Normalized profiles of TKE budget terms from the model grid point marked with A in Figure 2 at 12 UTC 7 July.

gradient even at the scalar wind-speed maximum. A consequence is that TKE shear production due to the cross-flow term helps to maintain the TKE in the jet. Also, the local friction velocity (2) does not drop to zero, as it does in Smeets et al. 1999. This facilitates local scaling, as otherwise normalized velocity variances tend to infinity in the wind-speed maximum.



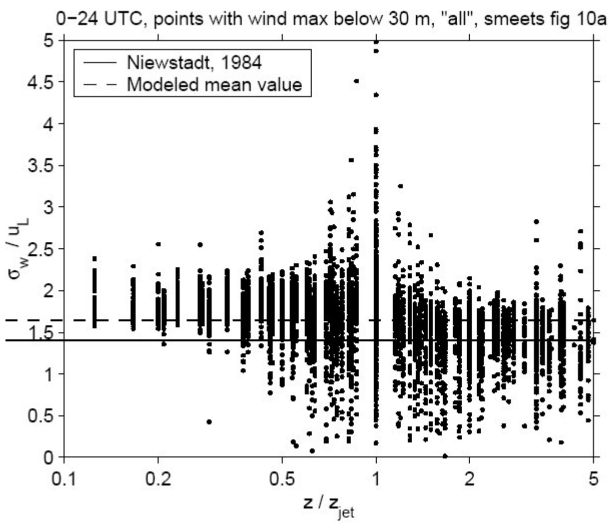


Figure 8 Standard deviation of the cross-wind (top) and vertical component (bottom) scaled with the local u_L plotted against height scaled with the height to the katabatic jet.

4. DISCUSSION

The boundary layer structure in a katabatic flow over a melting glacier has been investigated using numerical modeling. Among the findings are that many similarities to wind speed jets in other environments exist. Comparing for example to coastal marine jets, we find a similar vertical wave trapping and the turbulent flux of turbulence into the jets is also important in both. The significant departure of the local scaling results from the theoretical value in Niewstadt (1984) also behaves the same way as for the coastal jet in Brooks et al. (2003), although the dimensions of the jets themselves are very different. We hypothesize that this is due to a generality for boundary layers dominated by low-level jets, and moreover that the exact shape of this departure

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