# PARAMETERIZING SCALAR TRANSFER OVER A ROUGH ICE SURFACE

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## Background

Scalar transfer over ice surfaces in meso-scale and large-scale atmospheric models, and from Automatic Weather Station (AWS) data is typically calculated using the single level bulk-aerodynamic method. For this method the momentum ($z_0$) and scalar roughness ($z_s$) lengths are the key parameters. Usually $z_0$ in the ablation area during summer melt is taken to be constant while $z_s$ is calculated with a well known surface renewal model (Andreas, 1987). We test this method with data obtained in the ablation area of the Greenland ice sheet along the so-called K-transect (Figure 1).

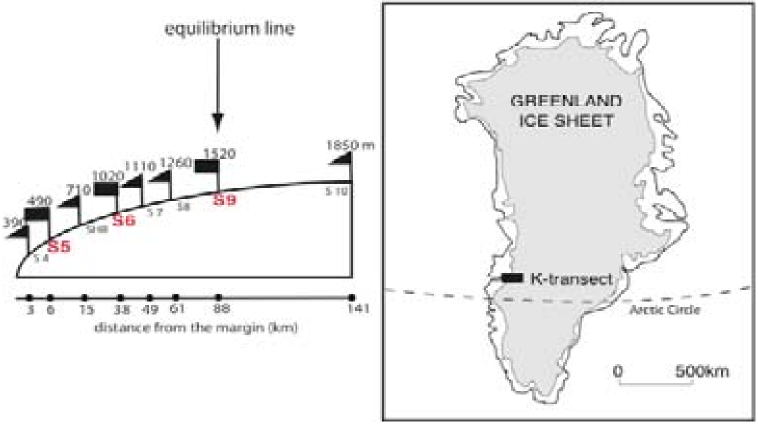

**Figure 1.** The K-transect and site locations in the ablation area of the Greenland ice sheet. AWS sites are S5, S6, and S9 (red).

Since August 2003 three Automatic Weather Stations (AWS) located at sites S5, S6 and S9 are operational (Figure 2). In addition, year-round Eddy-Correlation (EC) measurements were performed at S6 from August 2003 to August 2004 in order to study the surface processes.

### Bulk-aerodynamic method

The bulk-aerodynamic method relates, e.g. for sensible heat flux, the temperature difference between the surface and some height and a transfer coefficient with the turbulence flux:

$$H = \rho c_p C_{Hz} U(z)\left[T_0 - T(z)\right] \text{ and } C_{Hz} = \frac{k^2}{\left[\ln(z/z_0) - \alpha z/L\right]\left[\ln(z/z_T) - \alpha z/L\right]}$$

with $\rho$ the air density, $C_p$ the specific heat of air, $C_{Hz}$ the transfer coefficient, $U(z)$ the wind speed at height $z$, $T(z)$ temperature, $T_0$ the surface temperature, $k$ the von Karman constant (0.41), and $\alpha.z/L$ the static stability correction (relatively well known if the static stability is not large). The most uncertain parameters in this equation are $z_0$ and $z_T$ (subject of this abstract).

**Figure 2.** AWS (background) and eddy-correlation measurements (foreground) at S6.

## Momentum roughness lengths

The AWS are equipped with two measurement levels so that we can derive year-round $z_0$ estimates for S5, S6 and S9. However, good care has to be taken in the data analysis since $z_0$ derived from two level wind profile data is very sensitive to all kinds of errors. Thorough validation of AWS against EC data for location S6 resulted in strict selection criteria that guaranteed a selection of good quality $z_0$ results from our two level AWS data (see also Smeets and Van den Broeke, 2006). In Figure 3 we show the year-round results of $z_0$ for S5, S6, and S9 on the left, and a photograph from AWS5, the location with the roughest surface, at the end of the melt season in August 2003 on the right.

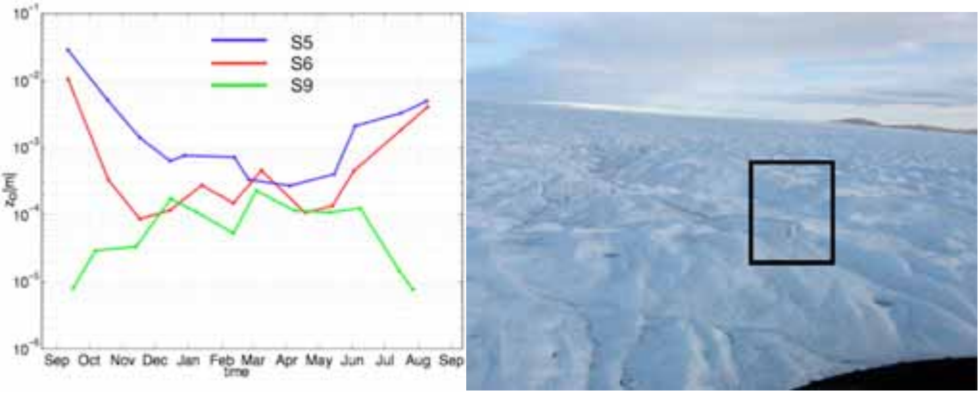

**Figure 3.** Left: year-round momentum roughness lengths ($z_0$) for location S5, S6, and S9. Right: a photograph of S5 taken from a helicopter at the end of the melt season in August 2003. The rectangle marks the AWS.

Throughout winter (November to May) the surface is in general snow covered and $z_0$ is fairly constant throughout the ablation area. However, when the melt season starts in May, spatial variations in $z_0$ rapidly increase to a factor 1000 between the lower ablation area (S5) and the equilibrium line (S9). These demonstrate the importance of large temporal and spatial variations of $z_0$ throughout a large part of the ablation area during the melt season. This result strongly contrasts with the general assumption in turbulence flux calculations in models or from AWS data of constant $z_0$ in the ablation area.

## Scalar roughness lengths

Scalar roughness lengths are usually calculated with the renewal model presented by Andreas (1987). Up to now the results of the model are validated over relatively smooth snow/ice surfaces. The data we are presenting here enables us to test the model results for a range of much larger $z_0$ values. First we straightforwardly compare the EC- and bulk-flux results. Bulk-fluxes are calculated with inclusion of a linear time dependence of $z_0$ as derived from our data. The surface temperatures are derived from longwave outgoing radiation measurements. We select cases with neutral to moderate stability. In Figure 4 we plotted the ratio of EC and bulk flux as a function of binned $z_0$ classes. We plotted sensible and latent heat flux results from S6 in Greenland. In addition, we plotted sensible heat fluxes from 2 locations obtained during an experiment on a broad outlet glacier at the Vatnajökull ice cap in Iceland in the summer of 1996 (e.g. Smeets et al., 1999).

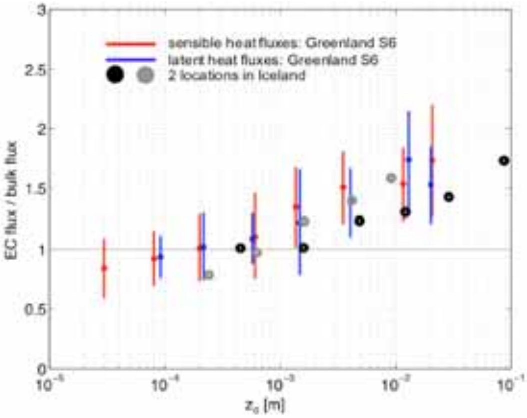

**Figure 4.** The ratio of EC and bulk-flux as a function of binned $z_0$ values. Year-round results from S6 on the K-transect in Greenland and from two locations on a broad outlet glacier from the Vatnajökull ice cap (Iceland, summer 1996).

In agreement with results from others, the fluxes agree well when the ice surface is flat and smooth with $z_0$ up to threshold values of 0.1 to 1 mm. However, when during the melt season the surface becomes hummocky, and $z_0$ grows substantially larger than the threshold, the ratio of EC- and bulk-flux increases well above 1 in a consistent fashion for all data (between 1.5 and 2 for $z_0 > 10$ mm).

We recognize that, in case of a flat ice surface, $z_0$ is dominated by surface stress originating from the substrate ice crystal cover. When the surface becomes hummocky, however, form drag rapidly starts to dominate $z_0$. Furthermore, our results indicate that a hummocky ice surface promotes substantially higher heat transfer then predicted by the model, meaning much too small $z_s$ values.

We hypothesize that the substrate ice crystal cover, that essentially controls the heat transfer close to the surface, is better 'ventilated' in case of a hummocky ice surface. Notice that this concept agrees with observations over rough vegetated surfaces that can have scalar transfer efficiencies comparable to momentum (e.g. Garratt, 1992). In other words, $z_0$ and $z_s$ can be equal in contrast to the usual findings in a rough flow regime that $z_0 >> z_s$.

As a convenient alternative to Andreas model for smooth ice/snow surfaces we fitted our hummocky ice data to the same model type yielding different coefficients: $\ln(z_T / z_0) = 3.5 - 0.7.\ln(R_*) - 0.1.\ln(R_*)^2$. In Figure 5 we plotted the results as is usually done in this business, that is $\log(z_s/z_0)$ versus $\log(R_*)$, with $R_* = u_* z_0/\nu$ the Reynolds roughness number. Both model curves are plotted together with our smooth and rough ice data. In addition, we plotted some results from vegetated surfaces (Garratt, 1992) to present our results in a broader perspective.

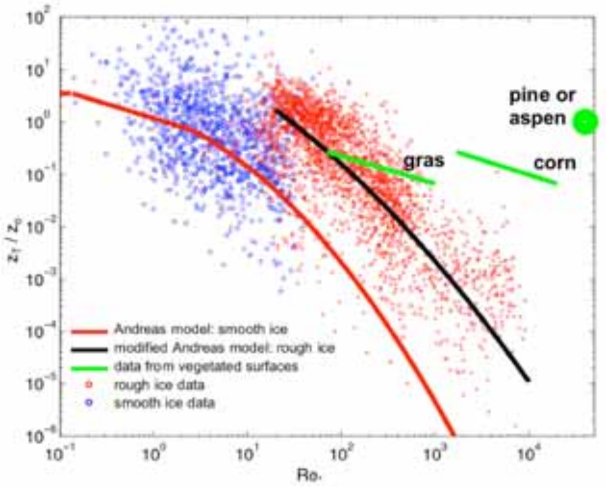

**Figure 5.** Ratio of scalar and momentum roughness lengths as a function of the Reynolds roughness number.

## Conclusions

Data from ablation areas in Greenland and Iceland show that the use of the bulk-aerodynamic method in models or for AWS data can be subject to large errors. The momentum ($z_0$) and scalar roughness ($z_s$) lengths are the key parameters and the former is usually taken constant while the latter is calculated with the model of Andreas (1987). The data illustrate that during the melt season, in strong contrast with assumptions, the momentum roughness lengths throughout the ablation area show very large spatial and temporal variations (between 0.1 and 10 cm). Furthermore, when the surface becomes hummocky during the melt season, the model of Andreas seriously underestimates heat transfer (up to 100%), i.e. the calculated scalar roughness lengths are too small. We suggest, in case of hummocky ice, the use of an alternative model fit derived from fitting our rough ice data.

## References

Andreas, 1987. A theory for the scalar roughness and the scalar coefficients over snow and ice. *Bound.-Layer Meteorol.,* **38**, 159-184.

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