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On the Friction Velocity during Blowing Snow

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Abstract

Snow concentration and wind profile measurements, which were gained during the Byrd Snow Drift Project in the Western Antarctic, are the basis of the investigation in the paper. It is shown that during episodes of snow drift, which is common feature in Antarctica, the influence of drifting snow particles on the density stratification cannot be neglected for studies of the turbulence parameters. A new approach is introduced by which these effects on the friction velocity can be described. It is shown that generally snow drifts near the surface lead to a significant stabilizing of the surface layer and a corresponding drag reduction. Inclusion of this effect results in an improved formulation of the snow drift densities. Comparisons between calculations and observations are presented.

Zusammenfassung

Die Schubspannungsgeschwindigkeit bei driftendem Schnee

Messungen der Schneedriftkonzentrationen und der Windprofile während des Byrd-Schneedrift-Experimentes in der westlichen Antarktis bilden die Basis der vorliegenden Untersuchungen. Es wird gezeigt, daß während Schneedriften, ein verbreitetes Phänomen in der Antarktis, der Einfluß der bodennah driftenden Partikeln auf die Dichteschichtung für die Beschreibung von Turbulenzparametern nicht vernachlässigbar ist. Es wird eine Methode dargestellt, mit der dieser Effekt auf die Reibungsgeschwindigkeit berücksichtigt werden kann. Bodennahe Schneedriften führen im allgemeinen zu einer signifikanten Stabilisierung der unteren Atmosphäre mit entsprechender Reduktion der Bodenreibung. Eine verbesserte Näherungsformel zur Berechnung der Schneedriftkonzentration wird abgeleitet und mit den Meßergebnissen verglichen.

1 Introduction

Mean wind and temperature profiles have been used for a couple of decades to derive turbulence quantities such as roughness lengths, friction velocities, sensible heat fluxes and stability parameters. These quantities are needed for the investigation of atmospheric processes within the surface layer and for modelling purposes. For nearly thermally neutral conditions, wind profiles are generally accepted to be logarithmic and procedures to derive above mentioned quantities are of relatively simple form. These approaches, however, require modifications in situations when particle transport takes place near the surface. Such phenomena are connected with sand and dust storms, sea spray, blowing snow episodes and with sedimentation processes in rivers and oceans. For the simulation of such situations, a two-phase flow together with stabilizing effects of the lower layers of the fluid have to be considered.

Within the region of particle transport the density stratification is determined by the particle concentration and by the thermal stability of the air. In many investigations, especially during snow drift, the influence of particles on the density stratification is generally neglected and only the effect of the thermal stratification is considered for studies of the turbulence parameters.

In this paper, we investigate snow drift episodes which were observed during the Byrd Snow Drift Project in Antarctica in 1962. A comprehensive description of the measurements and the data analysis is given by Budd et al. (1966) – in the following abbreviated as BDR. The relevant data are reproduced in the appendix of BDR. The pecularity of these observations under blowing snow conditions is the absence of any significant deviations from the logarithmic wind profile in the lowest 4 meters. However, we show that a formal application of the profile method to blowing snow conditions may lead to an overestimation of the surface friction velocity. With this aim we assume that air and drifting snow particles can be considered as a two-phase fluid. A simple theory of this phenomenon is presented in Section 2. The traditional profile method did not take into account the snow drift profile. Depending on available snow drift parameters, three variants of the modified profile method are proposed (Section 3) to estimate *diagnostically* the surface friction velocity. Comparisons between calculations and observations are presented in Section 4. Results of this study are summarized in Section 5.

2 Theory

The wind profile in the neutral atmospheric constant stress layer can be expressed by

$$u(z) = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right), \tag{1}$$

where $u(z)$ is the horizontal wind velocity at the level z, z_0 is the roughness length, u_* stands for the friction velocity and $\kappa = 0.4$ is von Karman's constant. The vertical wind profile is related to u_* by

$$u_* = \kappa m, \tag{2}$$

where $m = du/d(\ln z)$.

In stratified flows, stability effects under stable conditions and under weakly unstable conditions are taken into account according to Monin and Yaglom (1971) by

$$u_* = \frac{\bar{\rho}}{\bar{\rho}_0} \frac{\kappa m}{1 + \beta z/L}, \tag{3}$$

where ρ is the air density, ρ_0 is a value of ρ at the surface, $\beta > 0$ is a numerical coefficient, and L stands for the Monin-Obukhov length defined by

$$L = \frac{\bar{\rho} u_*^3}{\kappa g \overline{\rho' w'}} \tag{4}$$

Here w is the vertical wind velocity component, g is the acceleration of gravity and the quantities with bars and primes denote the statistical mean and turbulent fluctuations, respectively. The factor $\bar{\rho}/\bar{\rho}_0$ is usually neglected in Eq. (3).

Assuming that air and drifting snow particles can be considered as a two-phase fluid, the density ρ of the mixture may be expressed by

$$\rho = \rho_a(1 - S) + \rho_s S = \rho_a(1 + \sigma S), \tag{5}$$

where ρ_a and ρ_s are the respective densities of air and of snow particles, $S(z)$ is the volume concentration of snow particles and σ indicates the relative excess of the snow particle density over the air density:

$$\sigma = (\rho_s - \rho_a)/\rho_a$$

Assuming that

$$\bar{\rho} = \bar{\rho}_a(1 + \bar{\sigma}\bar{S}), \quad \bar{\sigma} = (\rho_s - \bar{\rho}_a)/\bar{\rho}_a, \tag{6}$$

and neglecting double and triple products of fluctuations, it follows from Eq. (5):

$$\rho' = \rho_a'(1 - \bar{S}) + \bar{\rho}_a \bar{\sigma} S',$$

and consequently the density flux $\overline{\rho' w'}$ in (4) may be expressed as follows:

$$\overline{\rho' w'} = \overline{\rho_a' w'}(1 - \bar{S}) + \bar{\rho}_a \bar{\sigma} \overline{S' w'} \tag{7}$$

Substitution of $\bar{\rho}$ from Eq. (6) and $\overline{\rho' w'}$ from Eq. (7) into Eq. (4) leads to the following expression for the stability parameter L:

$$L = \frac{\bar{\rho}_a(1 + \bar{\sigma}\bar{S}) u_*^3}{\kappa g (\overline{\rho_a' w'}(1 - \bar{S}) + \bar{\rho}_a \bar{\sigma} \overline{S' w'})} \tag{8}$$

The first term on right-hand side of Eq. (7) can be written in the following form when pressure fluctuations are negligible:

$$\overline{\rho' w'} = -\frac{\bar{\rho}_a}{\bar{\theta}} \overline{\theta' w'}, \tag{9}$$

where θ is the potential temperature. In the absence of snow particles ($\bar{S} \equiv 0$ and $\overline{S' w'} \equiv 0$) Eqs. (8) and (9) lead to the expression for the stability parameter L, derived by Monin and Obukhov (1954):

$$L = -\frac{\bar{\theta} u_*^3}{\kappa g \overline{\theta' w'}}.$$

In the case of thermally neutral stratification ($\overline{\theta' w'} \equiv 0$) the parameter L in the presence of drifting snow has the form:

$$L = \frac{(1 + s\bar{S}) u_*^3}{\kappa g \sigma \overline{S' w'}}, \tag{10}$$

and factor $\bar{\rho}/\bar{\rho}_0$ in Eq. (3) may be expressed by

$$\frac{\bar{\rho}}{\bar{\rho}_0} = \frac{1 + \sigma \bar{S}}{1 + \sigma \bar{S}_0}.$$

In this paper, we neglect thermal stratification of the air to study effects of the blowing snow on the friction velocity. For further simplicity in notation, the bar, signifying mean values, will be omitted everywhere except for the notation of the turbulent fluxes. We make an eddy diffusion assumption for the concentration flux of snow:

$$\overline{S'w'} = -K_s \frac{dS}{dz} \qquad (11)$$

and use the widely accepted balance relation of steady-state snow drift (BDR; Barenblatt and Golitsyn, 1974; Male, 1980; Soulsby and Wainwright, 1987):

$$K_s \frac{dS}{dz} = -w_s S, \qquad (12)$$

where K_s is the eddy diffusivity for snow and w_s is the falling velocity.

It was found from measurements (Baumeister and Marks, 1958; Fletcher, 1962; Budd, 1966) that the falling velocity w_s of snow particles of small diameter d (between 2 and 50 microns) is proportional to d^2, following the Stokes' law:

$$w_s = \frac{\sigma g d^2}{18\nu},$$

where ν is the molecular viscosity of air. For larger particles of $d > 2$ mm w_s is proportional to $d^{1/2}$, and for particles of intermediate sizes $0.1 < d < 1.0$ mm the falling velocity is proportional to d.

By integration of Eq. (12) over z, one can obtain from known parameters K_s and w_s the vertical distribution of the snow particle concentration S. If we take, as widely accepted for neutral conditions,

$$K_s = \kappa u_* z, \qquad (13)$$

and assume that w_s is constant with height, we then find the following well-known solution to Eq. (12):

$$S(z) = S_r (z/z_r)^{-\omega}, \qquad (14)$$

where $\omega = w_s/(\kappa u_*)$ and S_r is a prescribed reference value of S at height z_r.

It is seen from Eqs. (11) and (12) that the density flux is not constant with height during blowing snow and the Monin-Obukhov similarity theory cannot be applied in its traditional form. However, we can try to use the height-dependent density flux for the calculation of the stability parameter L.[1] This approach was used by Kondo et al. (1978) for the case of strong stability in the atmospheric surface layer and by Soulsby and Wainwright (1987) for the study of the effect of suspended sediment on near-bottom velocity profiles. The value of the parameter β in formula (3) will differ in this case from that found for scaling with the surface density flux. Kondo et al. (1978) found from observations of the atmospheric boundary layer that the most appropriate value of β for the height-dependent scaling equals 7, while a value of $\beta = 4.7$ was derived by Businger et al. (1971).

Taking into account Eqs. (10), (11) and (12), the following expression for the stability parameter L is valid:

$$L = \frac{(1+\sigma S)\, u_*^3}{\kappa g w_s \sigma S}. \qquad (15)$$

Since L is positive for the case of blowing snow, the value of u_*, derived from Eq. (2), will be an overestimate compared with the value of u_* given by Eq. (3).

If we account for stability effects, we may substitute for K_s in Eq. (12) by:

$$K_s = \frac{\kappa u_* z}{1 + \beta z/L}. \qquad (16)$$

Eqs. (12), (15) and (16) can be reduced to the following first-order differential equation:

$$\frac{dS}{dz} + \frac{w_s S}{\kappa u_* z}\left(1 + \frac{\beta \kappa g w_s z \sigma S}{(1+\sigma S)\, u_*^3}\right) = 0. \qquad (17)$$

Replacing $(1 + \sigma S)$ in this equation by the reference value $(1 + \sigma S_r)$, after some mathematics we then find the solution, subject to the boundary condition $S = S_r$ at a reference height $z = z_r$:[1]

$$S(z) = \frac{(1-\omega)\, S_r (z/z_r)^{-\omega}}{1 - \omega + \alpha\omega^2((z/z_r)^{(1-\omega)} - 1}, \qquad (18)$$

where

$$\alpha = \frac{\beta g \kappa^2 z_r \sigma S_r}{(1+\sigma S_r)\, u_*^2}.$$

For $\omega \to 1$ the solution (18) converges to:

$$S(z) = \frac{S_r (z/z_r)^{-1}}{1 + \alpha \ln (z/z_r)}. \qquad (19)$$

[1] A comprehensive review of similarity theory based on the hight-dependent stability parameter is presented by Stull (1988).

[1] For $\sigma S << 1$ the solution to Eq. (17) has been published by Taylor and Dyer (1977).

Since $(1 + \sigma S)$ varies with height, we use a semi-analytical approach to find solution to Eq. (12) with K_s given by formula (16). For this purpose we take $(1 + \sigma S)$ to be constant with height only between adjacent vertical levels of the snow concentration measurements and apply then formulae (18), (19). The lowest level of snow drift observations and the measured value of snow concentration z_1 and S_1 are used at first as reference values to calculate $S_2 = S(z_2)$ at the next level of measurements z_2. The resulting z_2 and S_2 values are used to derive in the same way $S_3 = S(z_3)$ etc. Owing to absence of the direct measurements of u_*, Eqs. (14) and (18), (19) will be subsequently applied to observations of the Byrd Snow Drift Project, summarized in BDR, to check effects of drifting snow on the friction velocity.

3 Modified Profile Method

Substitution of L from Eq. (15) into Eq. (3) leads to

$$\frac{1 + \sigma S}{1 + \sigma S_0} \kappa m = u_* \left(1 + \frac{\beta g \kappa w_s z \sigma S}{(1 + \sigma S) u_*^3} \right). \tag{20}$$

Let us consider the hypothetical case that S is given by formula (14) with $\omega = 1, \sigma S << 1$, and m assumed being constant with height. With these assumptions Eq. (20) is transferred into

$$\kappa m = u_* \left(1 + \frac{\beta g \kappa^2 z_r \sigma S_r}{u_*^2} \right). \tag{21}$$

It follows from Eq. (21), that in this special case u_* remains constant with height (the wind velocity profile is logarithmic, as empirical measurements from BDR indicate), but the presence of snow particles causes a decrease in the estimate of u_* compared to the situation without snow. We conclude therefore, that using profile method without respect to snow drift may lead to an overestimation of u_*. In general, the solution u_* in Eq. (20) can be not constant with height. However, if the vertical variations of calculated u_* are small enough with respect to its vertically averaged value, we may accept this value as the desired surface friction velocity.

If the snow concentration is known from measurements, the modified profile method can be applied to estimate the friction velocity. For this purpose we suggest three possible variants of this method.

Variant 1. Empirical functions are used to derive the falling velocity w_s on the basis of available particle size measurements.

Using this variant, we rewrite Eq. (20) in the following form:

$$u_*^3 - \frac{1 + \sigma S}{1 + \sigma S_0} \kappa m u_*^2 + \frac{\kappa \beta g w_s z \sigma S}{1 + \sigma S} = 0. \tag{22}$$

We are interested only in real solutions of this cubic equation. It is known from elementary algebra, that depending on coefficients, either Eq. (22) has one real (negative) root or all three roots of this equation are real (one negative and two positive). If positive solutions exist, we take the greatest root, since it is identical to Eq. (2) without snow drift. Generally speaking the friction velocity u_*, obtained with the help of Eq. (22), differs from that derived on the basis of Eq. (2), and for the same falling velocity w_s parameter $\omega = w_s/(\kappa u_*)$ will be different for Eq. (14) and Eqs. (18), (19), respectively. Eq. (22) has only the negative solution, if

$$\frac{\sigma S (1 + \sigma S_0)^2}{(1 + \sigma S)^3} > \frac{2}{27} \frac{\kappa^2 m^2}{\beta g w_s z}. \tag{23}$$

Variant 2. The snow concentration measurements are fitted with the help of Eq. (14) derive parameter ω for every observed vertical profile of S.

Using this modification of the profile method, we rewrite Eq. (20) to obtain the following form:

$$u_*^2 - \frac{1 + \sigma S}{1 + \sigma S_0} \kappa m u_* + \frac{\beta \kappa^2 g \omega z \sigma S}{1 + \sigma S} = 0.$$

The two solutions of this quadratic equation are

$$u_* = \kappa \left(\frac{1}{2} M \pm \sqrt{\frac{1}{4} M^2 - \frac{\beta g \omega z \sigma S}{1 + \sigma S}} \right), \tag{24}$$

where $M = (1 + \sigma S) m / (1 + \sigma S_0)$. We consider only the solution with the positive square root, since it is identical to Eq. (2) without snow drift. This solution does not exist, if

$$\frac{\sigma S (1 + \sigma S_0)^2}{(1 + \sigma S)^3} > \frac{1}{4} \frac{m^2}{\beta g \omega z}. \tag{25}$$

From Eq. (24) it is evident that u_* decreases with increasing snow concentration. Contrary to the variant 1, parameter ω here is the same for Eqs. (14) and (18), (19). With the value ω for every vertical profile of the snow concentration S and u_*, derived with the help of Eq. (2) or (24), we may calculate the falling velocities ($w_s = \kappa \omega u_*$) and compare them with empirical and theoretical estimations.

Taking into account the notation $m = du/d(\ln z)$, we may rewrite Eqs. (2) and (3) in the form:

$$(1 + \sigma S) K \frac{du}{dz} = (1 + \sigma S_0) u_*^2, \tag{26}$$

where the momentum eddy diffusivity K is given by Eqs. (13) and (16), respectively.[1] Eqs. (12) and (26) have the following integral solution:

$$\frac{S}{1+\sigma S} = \frac{S_r}{1+\sigma S_r} \exp\left(-\frac{u-u_r}{(1+\sigma S_0)\,\xi}\right), \qquad (27)$$

where $\xi = u_*^2/w_s$ and u_r is a reference value of u at height z_r.

Variant 3. The snow concentration measurements are fitted with the help of Eq. (27) to derive the parameter ξ for every observed vertical profile of S.

Using this variant, we rewrite Eq. (20) to obtain the following form:

$$u_* = \kappa\left(\frac{1+\sigma S}{1+\sigma S_0}\, m - \frac{\beta g z \sigma S}{(1+\sigma S)\,\xi}\right). \qquad (28)$$

From Eq. (28) it is also evident that u_* decreases with increasing snow concentration. This solution is negative, if

$$\frac{\sigma S(1+\sigma S_0)}{(1+\sigma S)^2} > \frac{m\xi}{\beta g z}. \qquad (29)$$

Similar to the variant 1, the parameter ω (= $u_*/(\kappa\xi)$) will be different, generally speaking, for Eqs. (14) and (18), (19). Moreover, the estimated values of the falling velocity w_s (= $\kappa\omega u_*$) may be also different.

Inequalities (23), (25) and (29) describe the range of appropriate input parameters, for which the suggested variants of the modified profile method can be applied to observed data to obtain the u_* under blowing snow conditions. These restrictions, if realized, reflect both shortcomings of the theory and incompleteness of the experimental data.

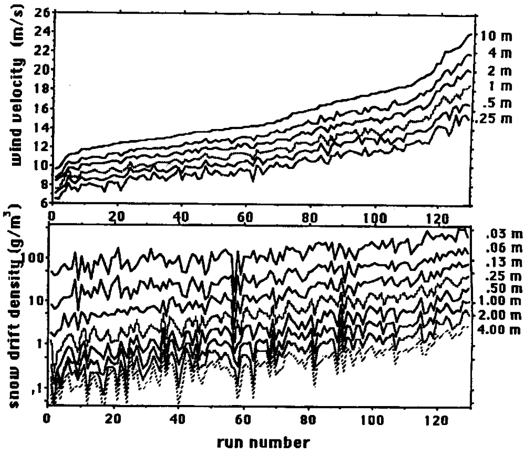

Figure 1 Byrd Station snow drift data. Upper part: Wind velocities, lower part: Snow drift densities. The data are sorted with regard to the wind velocity at 10 m. The numbers at the abzissa refer to different measuring episodes of 0.3 to 3.0 hours.

4 Calculations

To verify the effects of blowing snow, we apply measurements from the Byrd Snow Drift Project. These data include 129 episodes of 0.3 to 3.0 hours duration, during which snow traps were installed at 8 levels between 3.125 and 400 cm above the snow surface and simultaneous wind measurements were performed. Figure 1 shows the data which were sorted with regard to the wind velocity at 10 m. The snow drift data are given as drift densities n:

$$n = \rho_a \sigma S.$$

Table 1 Average wind velocity profile (m/s) for all the runs of measurements (N_m is the number of runs): observed (u_{obs}), logarithmic (u_{log}) and used in this study (u).

Height, m	0.03125	0.0625	0.125	0.25	0.50	1.0	2.0	4.0	10.0
N_m	4	61	14	127	119	105	127	129	129
u_{obs}	6.32	7.85	9.64	10.21	11.25	12.17	13.20	14.08	15.40
u_{log}	7.35	8.33	9.32	10.30	11.28	12.27	13.25	14.23	15.53
u	7.36	8.25	9.36	10.23	11.22	12.11	13.14	14.08	15.40

[1] Generally speaking, K for momentum and snow may be different, $K \neq K_s$, but any differences are neglected here.

To calculate σS the value $\rho_a = 1300$ g/m^3 was used. The wind velocity data (in BDR) were fitted with the help of Eq. (1), and values of u_* and z_0 derived. No evidence indicating any significant deviations of observed wind profiles form the logarithmic profile has been found. Table 1 confirms this observation. The logarithmic profile was applied to fill gaps in observations of the wind velocity.

To estimate the falling velocity w_s we used for all runs the mean of the particle size at different heights of measurements, given in BDR. It was found after vertical averaging that the particle size d can be estimated as equal to 0.0886 mm. Taking the empirical relation:

$$w_s = a d,$$

where $a = 2440\ s^{-1}$ (Budd, 1966), one can obtain $w_s = 0.216$ m/s. This value of w_s was subsequently used in calculations with variant 1 of the modified profile method.

The validity of the snow drift profiles with respect to Eq. (14) was tested. As a result, parameter $\omega = w_s/(\kappa u_*)$ has been derived for every snow drift episode. This information is an input for calculations with variant 2 of the modified profile method. Having ω and u_*, derived with the help of Eq. (2), one can obtain w_s for every run of measurements. These data are reproduced in the appendix of BDR. Using falling velocities from BDR, we have calculated average for all the runs w_s and found its value equal to 0.254 m/s, which is close to the above estimate.

The lowest level of snow drift measurements (3.125 cm) was used by us as the reference level. The reference value of the drift density was also taken to calculate the density of air-snow mixture at the surface. We have tested the validity of the snow drift profiles with respect to Eq. (27). As a result, parameter $\xi = u_*^2/w_s$ was derived for every snow drift episode and used in calculations with variant 3 of the modified profile method.

The observed snow drift densities and wind velocities were used as an input to estimate the friction velocity with the help of Eqs. (22), (24) or (28), respectively. Figure 2 displays the drift density n and the stability parameter L for the reference height (calculated using variant 1) as functions of the wind speed at 10 m height u_{10}. There is a significant increase of the drift density with wind speed for values above 10 m/s. The stability parameter L slightly decreases with wind speed from about 2–3 to about 1 m with a considerable scatter at lower wind values.

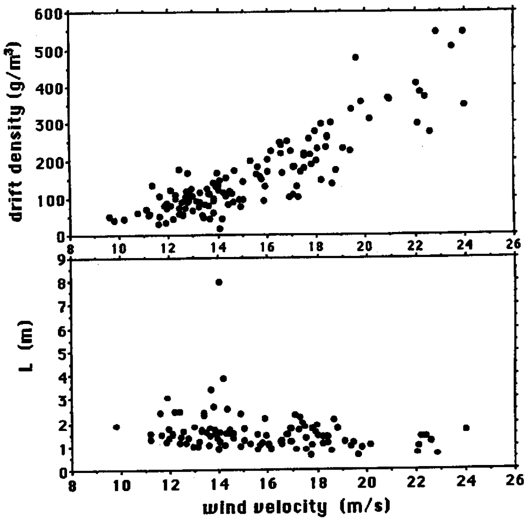

Figure 2 Byrd Station snow drift data. Snow drift densities and Monin-Obukhov stability parameter L (variant 1) for the height 3.125 cm as a function of the wind velocity (10 m).

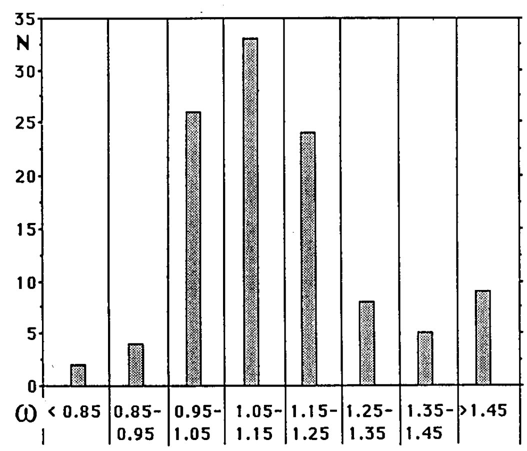

Figure 3 Histogramm of the parameter ω.

It should be noted that these values of L reflect strong stability with respect to the density of the air-snow fluid. Qualitatively the same behaviour of the stability parameter is also found in calculations with the other two variants of the modified profile method.

The histogram of ω-value in Figure 3 indicates that about 60 % of the data lie in the bins between 0.95 and 1.15. Consequently, the assumption of a logarithmic wind profile is still valid for these cases (see Eq. (21)), but a satisfactory determination of u_* requires the consideration of the snow drift parameters n and w_s.

An application of the modified profile method (variants 1, 2 and 3) was successful for 102, 111 and 129 snow drift episodes, respectively. The following ratio:

$$r = \Sigma/A,$$

has been applied to measure the degree of non-constancy of u_* with height. Here A and Σ are the vertical mean and standard deviation of every derived u_* profile. It was found that, in general, r did not exceed 5 %. Values of u_*, obtained with the aid of Eqs. (22), (24) or (28) for $\beta = 7$ and by Eq. (2) are plotted versus wind speed in Figures 4a–c. It is seen that, on the average, the first ones are distinctly lower than the second ones due to the stabilizing effect of snow drift. This effect is expressed most at wind velocity values above 18 m/s. Direct measurements of u_* have not been carried out during the Byrd Snow Drift Project, and thus we have no possibility to explicity test the validity of our concept.

In order to test this concept in an indirect way, we compared the drift densities obtained from Eqs. (14) and (18), (19) with the observed data. Results are presented in Figure 5a–c as an average for all the episodes. Although, a systematic difference still exists in variants 1 and 2 between the observed and calculated data, the consideration of the stability effect due to drifting snow achieves a remarkably better agreement between observations and theory.

Table 2 Mean friction velocities and snow drift parameters. N_e is the number of episodes, for which the modified profile method was successful, P.M. denotes the profile method based on Eq. (2), M.P.M.-k (k = 1, 2, 3) stands for the modified profile method of the k – th variant.

	u_* m/s	w_s m/s	ω	G g/m^2	Q g/(ms)	N_e
P.M.	0.586	0.216	0.967	48.84	689.88	
M.P.M.-1	0.481	0.216	1.146	18.44	216.75	102
observed				15.31	180.49	
P.M.	0.556	0.254	1.155	21.59	260.84	
M.P.M.-2	0.455	0.209	1.155	15.71	174.98	111
observed				14.63	175.97	
P.M.	0.567	0.250	1.073	25.11	275.39	
M.P.M.-3	0.471	0.170	0.901	13.53	164.78	129
observed				15.08	179.96	

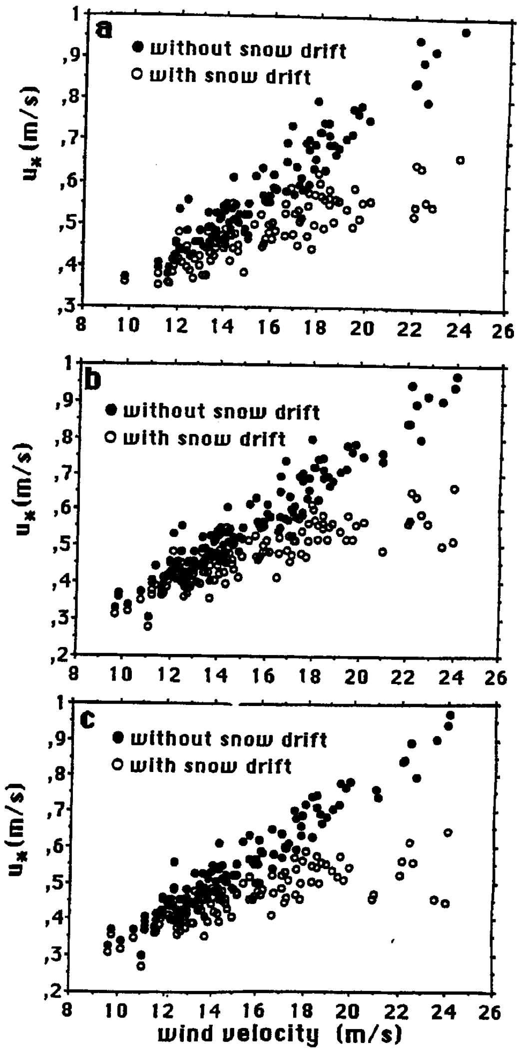

Figure 4 The dependence of the friction velocity on the wind velocity, calculated for situations with and without snow dirft: a) variant 1; b) variant 2; c) variant 3.

The same conclusion is true for the drift content G and for the drift transport rate Q, defined by:

$$G = \int_{z_r}^{h} n\,dz, \quad Q = \int_{z_r}^{h} un\,dz, \tag{30}$$

where h is the highest level of measurements (4 m in this study). Figure 6a–c shows the calculated and observed in different episodes drift contents which are sorted with regard to the wind velocity u_{10}. The plot for the snow drift transport rate Q is not

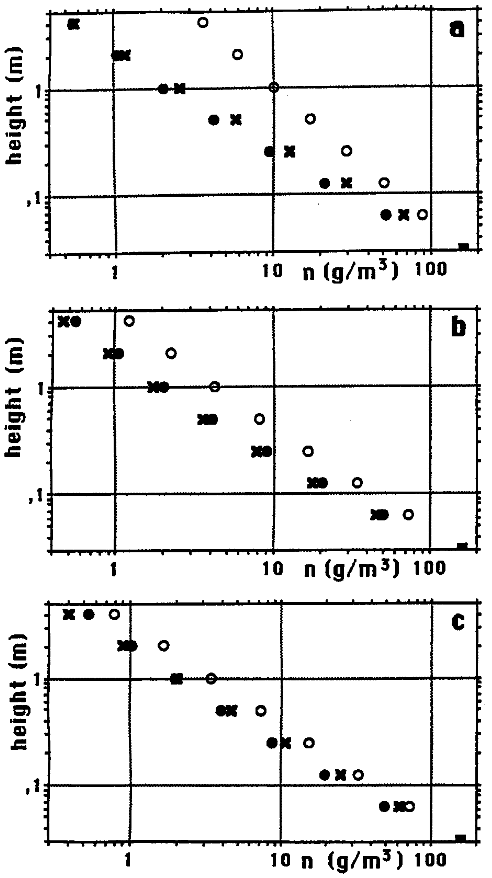

Figure 5 Observed and calculated snow drift densities versus height: a) variant 1; b) variant 2; c) variant 3. Filled circles are the observations, crosses and open circles are the calculations with and without regard of snow drift.

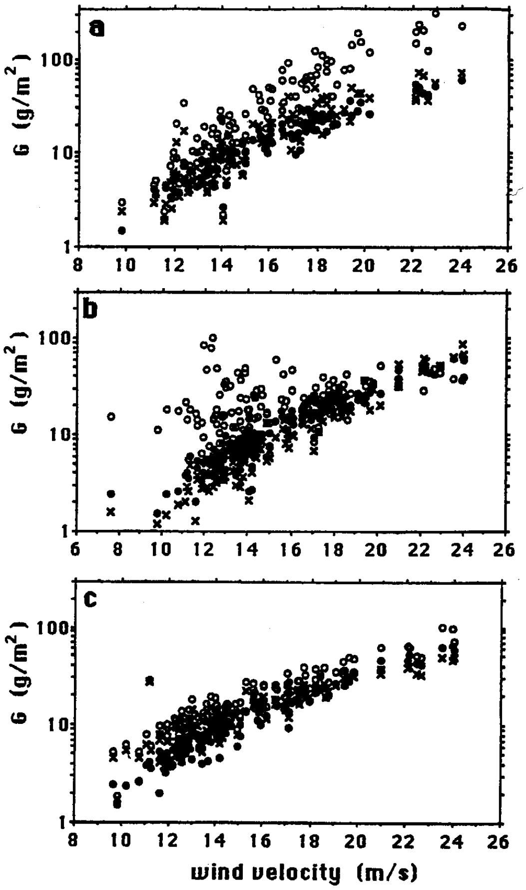

Figure 6 Observed and calculated snow drift contents versus wind speed at 10 m: a) variant 1; b) variant 2; c) variant 3. Filled circles are the observations, crosses and open circles are the calcualtions with and without regard of snow drift.

presented here since it shows a similar behaviour as for G.

Table 2 shows the friction velocity u_* and snow drift characteristics averaged for all the episodes, for which the modified profile method was successful. The table contains the falling velocity w_s, the parameter ω,the drift content G and the drift transport rate Q calculated with help of Eq. (2) and Eqs. (22), (24) and (28). Averaged drift content and drift transport rate are compared in Table 2 with the observed data. It seen from Table 2 that the modified profile method is not very sensitive to the choice of the version.

5 Discussion

Snow concentration and wind profile measurements, which were gained during the Byrd Snow Drift Project in the Western Antarctic, are the basis of the investigation in this paper. We have shown that during episodes of snow drift, which is a common feature in Antarctica, the influence of drifting snow particles on the density stratification cannot be neglected for studies of the turbulence parameters. The traditional profile method based on Eq. (2) did not take into account the snow drift profile to estimate *diagnostically* the surface friction velocity. A new approach is introducted by which these effects on the friction velocity can be described. Depending on available snow drift parameters, three variants of the modified profile method are proposed. It is shown that generally snow drifts near the surface lead to a significant stabilizing of the surface layer and a corresponding reduction of the surface friction velocity. We had no possibility to compare estimated friction velocities with direct turbulence measurements. However, inclusion of this effect resulted in an improved formulation of the snow drift densities. The conclusions, which we have drawn, are to some extent still qualitative. Direct observations of the wind velocity components together with accurate measurements of the snow drift density are needed to verify our approach.

If we are on the right track, then existing parameterization schemes of the atmospheric surface layer must be revised. We have calculated the 10 m drag coefficient C_d defined by:

$$C_d = \left(\frac{u_*}{u_{10}}\right)^2.$$

Values of C_d derived with the aid of traditional profile method and by the modified one (variant 1) are plotted versus u_{10} in Figure 7. It is seen that, on the average, while the first ones *increase* with increase of wind speed the second ones *decrease*. This agrees with variations of the stability parameter L (see Figure 3). Calculations with two other variants of the modified profile method also support this conclusion.

An approach, employed in this paper, can be also used for combined *prediction* of the wind velocity and snow drift profiles, when their reference values are prescribed, and for a possible explanation of quasi-logarithmity of the wind velocity profile under blowing snow conditions, as indicated by BDR. Assuming for the simplicity of analytical procedure $\sigma S \ll 1$, and substituting S form Eqs. (18), (19) into Eq. (26) with K given by Eq. (16), we obtain:

$$u(z) = u_r + \frac{u_*}{\kappa}\left(\ln\left(\frac{z}{z_r}\right) + \right.$$
$$\left. + \frac{1}{\omega}\ln\left(1 + \frac{\alpha\omega^2}{1-\omega}\left(\left(\frac{z}{z_r}\right)^{1-\omega} - 1\right)\right)\right). \qquad (31)$$

For $\omega \to 1$ the solution (31) converges to:

$$u(z) = u_r + \frac{u_*}{\kappa}\left(\ln\left(\frac{z}{z_r}\right) + \ln\left(1 + \alpha \ln\frac{z}{z_r}\right)\right). \qquad (32)$$

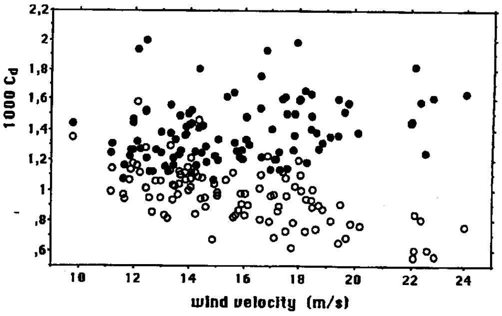

Figure 7 The dependence of drag coefficient on the wind velocity, calculated for situations with (variant 1) and without snow drift. Open circles and filled circles are the calculations with and without regard of snow drift.

If $z_r = z_0$, $u_r = 0$, and $\alpha = 0$, Eqs. (31), (32) coincide with Eq. (1). Comparing these equations, one can see that the snow drift effects on the wind velocity profile are also described by the additional logarithmic term. For $\omega \approx 1$ and $\alpha \ln(z/z_0) \ll 1$ Eq. (32) may be approximated by:

$$u(z) = u_r + \frac{(1+\alpha)\,u_*}{\kappa}\ln\left(\frac{z}{z_r}\right). \qquad (33)$$

Figure 3 and Table 2 show that, on the average, parameter ω, derived in BDR from the Byrd Snow Drift Project measurements, is close to 1. Eq. (33) demonstrates that the wind velocity profile will not differ in this case considerably from the logarithmic one. Table 1 supports this conclusion. Therefore, a formal application of the profile method on the basis of Eq. (1) to blowing snow conditions does not allow correct derivation of the friction velocity and the roughness length.

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