DEPOSITION AND EROSION
OF SNOW BY THE WIND

Uwe Radok

September 1968
PREFACE

This survey was prepared by Dr. Uwe Radok, University of Melbourne, working in the capacity of civilian expert for the Cold Regions Research and Engineering Laboratory (CRREL), U.S. Army Terrestrial Sciences Center (USA TSC). The author wishes to acknowledge the help received from Donald E. Nevel, James R. Hicks, Sherwood C. Reed, Austin Kovacs, and above all Malcolm Mellor, of USA TSC. Useful comments and suggestions were offered by William F. Budd in Melbourne.

USA TSC is a research activity of the Army Materiel Command.
CONTENTS

Preface .................................................................................................................. ii
Abstract ........................................................................................................ iv
Introduction ...................................................................................................... 1
Steady-state snow drift ....................................................................................... 2
Drift snow transport .......................................................................................... 3
The lower boundary ........................................................................................... 9
Saltation and suspension ......................................................................................
Obstacles and snow drift ..................................................................................... 12
Conclusions and suggestions ............................................................................. 20
Literature cited .................................................................................................... 21

ILLUSTRATIONS

Figure
1. Drift transport – wind speed relation for different layers ......................... 8
2. Drift sand concentration curves for Kurische Nehrung ............................... 13
3. Cumulative frequency curves of suspension up-currents for silt, snow and sand
   14
4. Mean fall velocities for Byrd Station snow drift particles ......................... 15
5. Major flow regions near obstacle (after Townsend) ..................................... 16
6. Major flow regions near obstacle (after Plate and Lin) ............................... 17
7a. Flow retardation ahead of obstacle ............................................................. 18
7b. Observed and computed drift heights upwind of Dye 3 ............................ 18

TABLES

Table
I. Snow drift transport as a function of the mean wind velocity $\overline{V}_{10}$ at the 10-m
   level .............................................................................................................. 5
II. Coefficients of polynomials in correction terms of eq 14 ............................ 7
III. Values of $D'$, $e^{kD'}$, and focal height $h$, as functions of 10-m wind speed ...
    11
The theories of uniform and non-uniform drifting snow are summarized with special emphasis on drift transport as a function of wind velocity. Using the work of Owen (1964) and the observations of the Byrd Station Snow Drift Project (Budd, Dingle and Radok, 1966) it is confirmed that the snow drift process involves a mobile surface layer of saltating particles, with a self-regulating thickness depending only on the surface stress and not on the snow concentration in the free air stream. It is shown to be a characteristic of snow (in contrast to sand or silt) that saltation and suspension drift occur side by side and that the latter reaches predominance as the wind velocity rises through the most common range of surface values. Theoretical reasons and observational evidence are produced for the view that deposition or erosion occurs on the snow surface during snow drift primarily as the result of mass flux convergence or divergence in the free air stream. This implies that the associated vertical mass flux penetrates the saltation layer which moves up or down with the snow surface. The survey concludes with suggestions for the experimental study of snow deposition and erosion in terms of the free air flow field and for a study of pneumatic particle transport in terms of saltation and of its electrical effects.
DEPOSITION AND EROSION OF SNOW BY THE WIND

by

Uwe Radok

INTRODUCTION

The effects of fluid flow on particulate matter in the fluid have been studied for many years, but only a small proportion of these studies has been concerned with gas-solid mixtures. The pioneer work in this field was Bagnold's (1941) book on blown sand which together with affiliated studies by Chepil (1945, 1946) and Zingg (1953) remains the standard reference for the case of large particle-fluid density differences and substantial free fall velocities. Bagnold's more recent efforts (1954, 1955, 1956, 1961) were aimed at a unifying set of laws for the entire practical range of density differences, but here a great deal remains to be done and even the characteristics of different regimes are as yet unknown. One such distinct regime appears to exist close to that of Bagnold's quartz sands and to produce different dominant factors for the wind behavior of snow, thereby making that material as distinctive in this context as in the thermodynamic and mechanical ones.

The subject of blowing snow has been reviewed by Mellor (1965) who gave special attention to studies concerned with physical causes and processes. Until recently most snow drift studies were concerned with practical aims such as the design of snow fences for clearing roads and buildings or for controlling snow characteristics in avalanche source regions. A change came with the IGY when in order to clarify the role of drifting snow in the Antarctic mass balance Mellor (1960) and Lister (1960) independently designed snow traps and subjected them to wind tunnel and field tests of collection efficiency. A little later Govorukha and Kirpichev (1961) described a snow trap of controlled collection efficiency which subsequently was redesigned by Landon-Smith for operational conditions (Landon-Smith and Radok, 1967). Thus the art of measuring fluxes of drifting snow was brought to the point where reliance could be placed on direct field measurements and where more advanced but indirect techniques could be utilized on an experimental basis (Landon-Smith, Woodberry and Wishart, 1965).

On the theoretical side basic turbulence concepts had been applied to the drifting snow phenomenon already by Schmidt (1925) but the fuller exploration of this approach may be dated from the independent work of Shiotani and Arai (1953) and Loewe (1956). Before that Diunin (1954) had attacked the problem by means of eddy models and dimensional reasoning but had to postulate, for realistic results, some details in need of explanation. Mellor and Radok (1960) initiated the experimental verification of the deductions made from turbulence theory and linked them with Bagnold's sand concepts. Extensive snow drift measurements in Antarctica by Dingle at Wilkes Station (Dingle and Radok, 1961) and especially at Byrd Station (Budd, Dingle and Radok, 1966) enabled Budd (1966) to give the first fully satisfactory explanation of the observed drift snow concentrations in terms of the observed momentum transport and drift particle size spectrum.

With this understanding of steady-state snow drift, attention must now be given to boundary and transient phenomena which not only create the most serious practical problems but also provide the link to more general questions of fluid particle transport. The specific problem discussed in this survey concerns the nature of the factors controlling the wind deposition and erosion of snow. As a start the present knowledge of steady-state snow drift will be summarized.
A detailed discussion of the laws governing the drifting of snow some distance above the snow-air boundary is given in the reports on the Byrd Snow Drift Project (Budd, Dingle and Radok, 1966; Budd, 1966). Here it will suffice to state the results of this work and the extent to which they are supported by observational evidence.

The concentration, \( n_z \), at height \( z \) of a mass of uniform snow particles with fall velocity \( w \) is governed by the relation

\[
n_z w = k \left( \frac{u^*}{z} \right) \frac{\partial n_z}{\partial z}
\]

where \( k \) and \( u^* = (r/\rho)^{1/2} \) are the von Karman constant and shear velocity in the logarithmic wind profile \( V_z = (u^*/k) \log_e (z/z_0) \), valid for a surface roughness length \( z_0 \) and the constant stress \( (r) \) layer in the realistic conditions of neutral stratification. Integration of eq 1 gives the drift concentration relative to that at a reference level \( Z \) as

\[
n_z = n_z \left( \frac{z}{Z} \right)^{-w^*}
\]

where \( w^* = w/ku^* \). Individual drift concentration profiles, derived from parallel drift snow collections at different levels for periods ranging from 5 minutes to half an hour, confirm eq 2 as a reasonable approximation.

Formally the logarithmic wind profile makes it possible to express \( n_z \) as a function of wind velocity, since \( u^* = V_z.k/\log_e(z/z_0) \); thus eq 2 becomes

\[
n_z = n_z \exp \left[ \frac{-w}{k^2} \log_e \left( \frac{z}{z_0} \right) \right] \log_e \left( \frac{z}{z^2} \right) V_z^{-1}
\]

Observations show that eq 3 is quite well realized when \( n_z \) is taken at a low level where it is found to be substantially independent of wind speed. The implications will be discussed later.

Closer study of a large number of vertical drift concentration profiles reveals that although eq 2 is valid for snow particles of a given size (and hence a given fall velocity \( w \)) it does not describe the average features of non-uniform snow drift except near the surface or for large wind speeds. An improved theory to account for the discrepancies was devised by Budd (1966) who approximated the observed distributions of drift snow particle diameters \( \xi_z \) with gamma distributions of the form

\[
\text{Prob} \left( x < \xi_z < x + dx \right) = \frac{1}{\beta^\alpha \Gamma(-\alpha)} e^{-x/\beta} x^{\alpha-1} \ dx
\]

where the parameters \( \alpha \) and \( \beta \) are defined by \( a\beta = \xi_z \) and \( a\beta^2 = \xi_z^2 \). In the observed drift particle size range (30 to 500\( \mu \)) the fall velocity is represented by \( w = \xi/c \), where \( c \) is a constant of the order of \( 3 \times 10^{-2} \) sec. Budd then showed that the concentration of non-uniform drift snow at height \( z \) is given by

\[
N_z = n_z \left[ 1 + \left( \frac{\beta_z}{cku^*} \right) \log_e \left( \frac{z}{z} \right) \right]^{-aZ+3}
\]

This relation again can be readily expressed in terms of wind velocity and is in very good agreement with the observations, even in the mean. As byproducts of his analysis Budd obtained the
expressions for the particle mean diameter, fall velocity, and number per unit volume, as functions of height and wind speed ($u^*$):

Mean particle diameter

$$\bar{x}_Z = \bar{x}_Z \left(1 + \frac{\beta_Z}{ku^* \log e \frac{Z}{Z}}\right)^{-1}. \quad (6)$$

Mean fall velocity

$$\bar{w}_Z = \bar{w}_Z \left(1 + \frac{\beta_Z}{cu^* \log e \frac{Z}{Z}}\right) = \frac{(aZ+3)ku^*}{\log e \frac{Z}{Z}} \left[\left(\frac{N_Z}{N_Z}\right)^{1.3} - 1\right]. \quad (7)$$

Particle number per unit volume

$$\nu_Z = \nu_Z \left(1 + \frac{\beta_Z}{cu^* \log e \frac{Z}{Z}}\right)^{-aZ} \quad (8)$$

where

$$\nu_Z = \frac{N_Z}{\rho \beta_Z^3 n Z} \frac{\Gamma(aZ)}{\Gamma(aZ+3)} = 2N_Z \text{ cm}^{-3} \ (N_Z \text{ in g m}^{-3}).$$

The approximate form of $\nu_Z$ arises from the fact that the parameter $a$ has a value close to 15 for observed snow drift particle size distribution, so that $\Gamma(a+3)/\Gamma(a) = (15 \times 16 \times 17) = 4080$; this value of $\nu$ is close to the number of uniform particles of 0.1 mm diameter representing a concentration of $n_Z$ g m$^{-3}$.

The restriction implied by the assumption of a gamma-shape size spectrum is not serious since in practice the gamma distribution cannot be distinguished easily from the log normal distribution characterizing fragmentation processes. Formulas 6 to 8 and the frequency distributions from which they were derived have been shown by Budd (1966) to be in good agreement with the Byrd Station observations. In addition, eq 8 was used successfully by Landon-Smith and Woodberry (1965) to account theoretically for the calibration of their photoelectric snow drift gauge.

Both in general and especially in the present context crucial importance attaches to the horizontal snow drift transport rate. For its computation the assumption of uniform drift was found to be inadequate (Budd, Dingle and Radok, 1966), and therefore the first problem to be considered here will be the implications of Budd’s (1966) non-uniform drift theory for drift transport.

**DRIFT SNOW TRANSPORT**

The drift transport through a vertical surface of unit width perpendicular to the wind direction is given formally by

$$q_Z = \int Z V_Z n_Z \, dz \quad \text{for uniform snow} \quad (9)$$
DEPOSITION AND EROSION OF SNOW BY THE WIND

or

\[ Q_Z^Z = \int_{z}^{Z} V_z^N z^N dz \quad \text{for non-uniform snow.} \quad (9') \]

The first of these integrals was evaluated by Dingle and Radok (1961) and for \( n_Z^* \) constant and \( w^* \neq 1 \) (a trivial restriction) can be written

\[ Q_Z^Z = \left( \frac{u^*}{1-w^*} \log e \frac{(z/z_0)^{zn_Z^*}}{(Z/z_0)^{zn_Z^*}} - \frac{u^*}{(1-w^*)^2} \right) (zn_Z^* - z n_Z^*) \quad (10) \]

However, examination of the Byrd Station data shows that eq 10 gives values substantially in excess of those obtained by adding up the transports in a number of layers of differing fall velocity. The discrepancy can be traced back to the change in mean fall velocity which occurs as the increasing wind speed draws heavier and heavier snow particles into the drift process. Table I gives the observed mean fall velocities for the Byrd Station data grouped into six wind speed groups and compares the transports computed from these fall velocities with those based on the overall mean fall velocity and with the sums of the layer transports.

The transport figures in column 6 of Table I make allowance for the change in fall velocity with both wind speed and height. If the change with height is neglected a slight upward bias is introduced by the substitution of a straight density profile for the actual curved one (column 5). This bias suggests the use of a slightly enlarged effective fall velocity. Comparison with the transport figures computed on the basis of the overall mean velocity shows that the change in particle size with wind speed is the principal factor to be taken into account, whereas the change in the particle size spectrum with height introduces only relatively minor adjustments to the transport figures.

The theory of non-uniform snow drift (Budd, 1966) has been found to account satisfactorily for the variations in the observed particle mean fall velocity with both wind speed and height. This suggests that a completely adequate drift transport formula might be derived from the integral 9'.

Substitution of \( \log e z = x \) in the full form of eq 9'

\[ Q_Z^Z = \int_{z}^{Z} u^* \log e \frac{z}{z_0} n_Z^* (1 - \frac{\beta x}{c k u^*}) \log e \frac{z}{Z} \left( \frac{(a x^3)}{1 - \frac{\beta x}{c k u^*}} \right) \quad (9'') \]

gives

\[ Q_Z^Z = \int_{x}^{X} V_x N_x e^x dx = N_X^* \frac{u^*}{k} \int_{x}^{X} e^x (x-x_0) \left( 1 - \frac{\beta x}{c k u^*} \right) \left( \frac{(a x^3)}{1 - \frac{\beta x}{c k u^*}} \right) \quad (9'''') \]

which has the form of

\[ A \int e^x [1 - c_1 (x - c_2)]^{-C_3} dx - B \int e^x [1 - c_1 (x - c_2)]^{-C_3} dx \]

exponential integrals without closed-form solutions. However the exponent \( -(a x^3) \) in eq 9''' and 9''''' is a large number and the first term or terms of a suitable series expansion should therefore suffice to represent \( Q_Z^Z \).
Table I. Snow drift transport as a function of the mean wind velocity $V_{10}$ at the 10-m level.

Height range: 1 mm – 300 m. Figures are based on the data of the Byrd Drift Project
(Budd, Dingle and Radok, 1966).

<table>
<thead>
<tr>
<th>Mean wind velocity $V_{10}$ (m sec$^{-1}$)</th>
<th>Drift transport</th>
<th>Corresponding transport</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>No. of runs</td>
<td>$\sigma_{10}^{-1}$ for fall vel.</td>
</tr>
<tr>
<td>-----------------------------------------</td>
<td>-------------</td>
<td>---------------------------------</td>
</tr>
<tr>
<td>11.5</td>
<td>25</td>
<td>71.9</td>
</tr>
<tr>
<td>12.7</td>
<td>25</td>
<td>128</td>
</tr>
<tr>
<td>14.5</td>
<td>25</td>
<td>314</td>
</tr>
<tr>
<td>16.8</td>
<td>25</td>
<td>1098</td>
</tr>
<tr>
<td>19.5</td>
<td>15</td>
<td>4198</td>
</tr>
<tr>
<td>23.4</td>
<td>14</td>
<td>20943</td>
</tr>
</tbody>
</table>

One such expansion was used by Budd (1966) to demonstrate the essential difference between the drift snow concentration formulas (eqs 2 and 5). Expansion of the logarithm of eq 5 gives

$$\log_{e} \frac{N_{Z}}{N_{Z}} = -w_{Z}^* \log_{e} \frac{Z}{Z} + \frac{1}{2} w_{Z}^* \log_{e} \frac{2}{Z} + \ldots$$

where $w_{Z}^* = (aZ + 3) \beta Z/c$, $w_{Z}^* = (3 \beta Z/c)/k u^*$. By contrast we obtain from eq 2

$$\log_{e} \frac{n_{Z}}{n_{Z}} = -w^* \log_{e} \frac{Z}{Z} + \ldots$$

Thus the non-uniformity of the drift snow produces a slightly larger effective mean fall velocity $w^* = 3 \beta Z/c$ in addition to an infinite series of positive correction terms. The effective fall velocity represents that of the particle of mean mass and is 20% larger than the mean fall velocity.

Using the first and second order terms in eq 11 only the drift transport takes the form

$$Q_{X} = \int_{X}^{X} N_{X} V_{X} e^{x} dx \exp \left(\frac{1}{2}(w_{Z}^* - w^*) \frac{X}{X} \right) N_{X} \frac{u^{*}}{k} \int_{X}^{X} x \exp \left(\frac{1}{2}(1 - w_{Z}^* - w^*) \frac{X}{X} \right) \sqrt{\frac{w_{Z}^*}{2}} x \frac{2}{dx}$$

where again $x = \log_{e} z$. The integrals are of the form

$$A \int x e^{c x^2} dx + B \int e^{c x^2} dx$$
DEPOSITION AND EROSION OF SNOW BY THE WIND

and can be reduced to error functions with complex argument which present similar difficulties as before.

We finally try a new approach leading to an additive correction for the uniform drift density profile (eq 2), in place of the multiplicative one implied by the power series (eq 11). Equation 2 as it stands can be expanded as follows:

\[ n_Z = n_Z \left( \frac{Z}{Z} \right)^{-w^*} = n_Z \exp \left( -w^* \log_e \left( \frac{Z}{Z} \right) \right) \]

\[ = n_Z \left[ 1 - w^* \log_e \left( \frac{Z}{Z} \right) + \frac{1}{2} w^*^2 \log_e^2 \left( \frac{Z}{Z} \right) + \ldots \right]. \tag{13} \]

The corresponding expansion for the non-uniform drift density (eq 5) is

\[ N_Z = N_Z \left[ 1 + \frac{\beta_Z}{c k u^*} \log_e \left( \frac{Z}{Z} \right) \right]^{-\left( a_Z + 3 \right)} \]

\[ = N_Z \left[ 1 - (a_Z + 3) \frac{\beta_Z}{c k u^*} \log_e \left( \frac{Z}{Z} \right) + \frac{(a_Z + 3)(a_Z + 4)\beta_Z^2}{2(c k u^*)^2} \log_e^2 \left( \frac{Z}{Z} \right) + \ldots \right] \tag{14} \]

Comparison of eq 14 and 13 confirms the previous conclusion that in computing drift transports the effective fall velocity should be taken as \( w_Z = w_Z + 3\beta_Z/c \). The correction terms now have the form

\[ \frac{1}{n!} P_n(a_Z + 3) \frac{\beta^n}{(c k u^*)^n} \log_e^n \left( \frac{Z}{Z} \right) \]

where \( P_n(a_Z + 3) = p_{n-1}(a_Z + 3)^{n-1} + p_{n-2}(a_Z + 3)^{n-2} + \ldots + p_1(a_Z + 3) \). The values of the coefficients \( p \) are given for the first few powers in Table II.

The form of the expression \( [(\beta_Z/c k u^*) \log_e (z/Z)] \) reflects the observed trend to greater uniformity with increasing wind speed. To estimate its magnitude we note that \( a_Z\beta_Z/c = \bar{w}_Z \), the mean fall velocity. For \( a_Z \) the Byrd Station data suggested values of the order of 15. Thus

\[ 0 \left( \frac{\beta_Z}{c k u^*} \log_e \frac{Z}{Z} \right) = \frac{\bar{w}_*}{15} \log_e \frac{Z}{Z} = \frac{1}{15} \log_e \frac{Z}{Z} \]

since \( w_* \) does not differ greatly from unity. The coefficients in Table II can then be used to estimate the approximate height ratios \( z/Z \) for which each term of the series of corrections remains one order of magnitude smaller than the preceding one.

The results of this calculation are given in the last three columns of the table. They show clearly that the slow convergence of the series of correction terms makes it useless for the present purpose of finding a simple drift transport formula. In fact the second order term is smaller than the third order term and hence quite useless except in the immediate neighborhood of the reference level. But even for this term the additive correction to the uniform drift transport formula gives rise to a cumbersome height integral.
Table II. Coefficients of polynomials in correction terms of eq 14.

<table>
<thead>
<tr>
<th>n</th>
<th>p_6</th>
<th>p_5</th>
<th>p_4</th>
<th>p_3</th>
<th>p_2</th>
<th>p_1</th>
<th>( C = \frac{P_n(18)}{n! 15^n} )</th>
<th>( \frac{C_n}{C_{n-1}} )</th>
<th>( \left( \frac{z}{Z} \right)<em>{\text{max}} = \exp(10C_n/C</em>{n-1})^{-1} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td></td>
<td></td>
<td>1</td>
<td>-3</td>
<td>-2</td>
<td>-0.0498</td>
<td>-1.244</td>
<td>1.084</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>-10</td>
<td>85</td>
<td>215</td>
<td>274</td>
<td>129</td>
<td>0.0318</td>
<td>-0.438</td>
<td>1.257</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>15</td>
<td>85</td>
<td>215</td>
<td>274</td>
<td>129</td>
<td>-0.0199</td>
<td>-0.337</td>
<td>1.345</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>-21</td>
<td>-175</td>
<td>-725</td>
<td>-1564</td>
<td>-720</td>
<td>0.0047</td>
<td>-0.278</td>
<td>1.433</td>
<td></td>
</tr>
</tbody>
</table>

\[
I(z) = \int \frac{\log e}{Z} \frac{z}{z_0} \log e \frac{z}{Z} dz
\]

\[
= \int \left[ \log e \frac{3z}{Z} - \log e \frac{2z}{Z} (2 \log e Z + \log z_0) + \log z (2 \log Z \log z_0 + \log^2 Z) - \log Z \log z_0 \right] dz
\]

which must be evaluated very accurately to avoid large errors in the final result.† The higher order terms involve even more elaborate expressions so that the numerical work involved in the use of additive correction terms would rapidly approach or exceed that of evaluating the full integral (eq 9″″ or 9″″″).

It appears therefore that the way to a simple analytical drift transport expression is barred on all sides, and that for practical purposes empirical relationships such as that given by Budd, Dingle and Radok (1966) for the drift transport at Byrd Station must be used instead. The Byrd Station relationship for the layer between the levels 1 mm and 300 m above the snow surface

\[
\log Q_{10}^{300} = 1.1812 + 0.0887 V_{10} (Q \text{ in g m}^{-1} \text{ sec}^{-1}, \ V_{10} = 10\text{-m wind speed in m sec}^{-1})
\]

† The evaluation of the integral was carried out under the direction of Mr. D.E. Nevel of USA CRREL on the CRREL computer.
DEPOSITION AND EROSION OF SNOW BY THE WIND

Figure 1. Drift transport – wind speed relation for different layers

(Byrd Snow Drift Project data)

Estimated drift transported in different layers.

Measurements of drift at a Byrd Station study site indicate that the percentage of drift at Byrd Station is attributed to incoming snowfall, and that the drift contribution is not affected by the wind speed. The drift contribution does not reveal the changing contributions made by different layers as the wind speed increases. These become important for the case of drift formation near obstacles and have therefore been prepared, from the Byrd Station drift data, to demonstrate the differing behavior of the drift transports in the surface and upper layers, respectively, given approximately by

\[
\log Q^{0.125}_{10^{-1}} = 1.36 + 0.0545 V_{10}
\]

and

\[
\log Q^{300}_{2} = 0.255 + 0.127 V_{10}
\]
The intermediate layer from 12.5 cm to 2 m in which most of the actual measurements were made does not contribute significantly to the total transport. The transport figures for this layer have however been included in Figure 1 to demonstrate their tendency to parallel the surface layer transports for high wind speeds and those of the upper layer for low wind speeds. They also serve to underline the largely extrapolated nature of the total drift transport estimates and the need for actual drift density measurements, by optical or other indirect techniques, at levels well above and below those so far studied.

THE LOWER BOUNDARY

All the formulas presented in the preceding two sections include the drift concentration \( n_z \) or \( N_z \) (for non-uniform snow) at a reference level \( Z \). For practical purposes this level must be chosen close to the snow surface where the heaviest drift takes place yet where direct measurements are difficult. The surface very often represents the only source of the drift snow and invariably supplies a substantial proportion of it, even though it also forms the drift snow "sink" in places where deposition occurs.

The difficulties of observing close to the surface lend special interest to any conclusions, regarding surface processes, which can be drawn from measurements higher up. The available evidence permits a number of such conclusions.

In the first place the extrapolation downward of both wind and drift snow concentration profiles for different wind speeds at a reference level (usually 10 m) suggests the existence of crossover points near the 1-cm level. This implies that different drift regimes prevail above and below that level, a fact already suggested strongly by the existence of a distinctive form of low-level or "creeping" drift.

A second and quite independent piece of evidence is provided by the observed wind dependence of the drift concentration at a given level. According to eq 3 this should have the form of a straight line in a \( \log n_z \) versus \( V_z^{-1} \) coordinate system provided \( n_z \) were independent of \( V_z \). The Byrd Station observations (Budd, Dingle and Radok, 1966, Fig. 26) actually show the drift concentrations to behave in this manner, with random scatter suggesting changes in surface drift concentration linked to existing meteorological conditions and perhaps different snow cover histories. Extrapolation of the curves for different levels to very large wind velocities indicates a trend towards a vertically uniform drift concentration similar in magnitude to the air density, \( 10^3 \text{ g m}^{-3} \); and since the rate of the increase in drift concentration with wind velocity decreases with decreasing height it is reasonable to conclude that the surface drift concentration will be of the same order of magnitude regardless of wind speed. Any adequate model of the surface drift process ought to contain this quantitative feature.

The most probable model for the surface layer drift is provided by Bagnold's "saltation," which was adapted to snow by Mellor and Radok (1960) and explored in numerical trajectory calculations by Jenssen (1963). However for the present purpose the greatest interest attaches to a significant theoretical extension of Bagnold's work which has recently been published by Owen (1964).

In Owen's study the saltation process is visualized as starting when hydrodynamic forces lift loose particles from the surface and in collaboration with gravity return them there with sufficient speed to dislodge other particles. Extensive observations have shown that this process sets in when the non-dimensional ratio \( \gamma = r/\rho g \xi \) reaches a threshold value \( \gamma_s \). For the sand and solid particles regime this threshold occurs at a value of \( \gamma_s = 0.0064 \) and it is of interest to examine its implications for snow particles. With \( \rho = 0.9 \text{ g cm}^{-3} \) and \( \xi = 0.1 \text{ mm} \) the threshold shear stress becomes \( \tau_s = 0.0064 \rho \gamma_s u_{*}^2 \), where \( \delta = 1.25 \times 10^{-6} \text{ g cm}^{-3} \) is the air density. This makes \( u_{*} = 6.7 \text{ cm sec}^{-1} \) and under the conditions at Byrd Station (where \( u^* = 26.5V_{10} \), cf. Budd, Dingle and Radok, 1966) corresponds to a 10-m wind velocity of 1.8 m sec\(^{-1}\) for the onset of surface drift.
This velocity is of the right order of magnitude. Kungurtsev (1956) gives a range from 4 m sec\(^{-1}\) down to 2 m sec\(^{-1}\) ("in special circumstances") at the 2-m level where the wind velocity is 92\% of that at the 10-m level, for average values of the roughness length \(z_0\). The Owen theory does not require a precise value of \(\gamma\) to be exceeded but merely that this parameter should lie in the range from \(10^{-2}\) to 1. The upper limit (which for snow corresponds to \(V_{10} = 22\) m sec\(^{-1}\)) will be critically examined in the next section and is not relevant in the present context.

The mechanism of saltation is explained by Owen (1964) in terms of two major working hypotheses. The first of these is that the saltation layer affects the air flow above it as would fixed surface roughness elements of comparable height: this can be verified by means of wind profiles, and Owen shows that Bagnold's (1941) winds support this hypothesis. The second hypothesis is that for a given boundary layer stress in the saltation range defined by \(10^{-2} < \gamma < 1\) the saltation layer is self-maintaining: a small increase in stress will produce momentarily more saltating particles so that by transfer of extra momentum to the surface layer the previous conditions will be restored. The second hypothesis leads to a relation for the saltation transport as proportional to \(u^*\) (in accordance with the findings of Bagnold and others) as well as to so far untested conclusions regarding the particle concentration in the saltation layer. Both these hypotheses will now be examined in the light of the Byrd Station snow drift data.

The vertical velocities to be expected in a flow of shear velocity \(u^*\) are themselves of the order of \(u^*\) (for a more detailed specification cf. next section) and hence the saltation layer thickness must be of the order of \(u^*/2g\), the maximum height that can be attained by a particle leaving the surface with the perpendicular velocity \(u^*\). The logarithmic wind profile \(V_z = (u^*/k) \log_e(z/Z)\) then takes the form (Owen, 1964)

\[
\frac{V_z}{u^*} = k^+ \log_e\left(\frac{2gz}{u^*}\right) + D'.
\]

\(D'\) is proportional to the logarithmic ratio of the saltation layer height to the usual roughness length \(z_0\); this can be seen by combining the standard form of the logarithmic wind profile with Owen's (eq 15) whence

\[
kD' = \log_e\left(\frac{u^*^2}{2g}\right) - \log_e z_0. \tag{16}
\]

From the Byrd Station measurements (Budd, Dingle and Radok, 1966) the following relations were found for the shear velocity and roughness length as functions of the 10-m wind speed:

\[
u^* = \frac{V_{10}}{26.5} \text{ m sec}^{-1}
\]

\[
\log_{10} z_0 = 1.104 + 0.039 V_{10} \text{ (in m, } V_{10} \text{ in m sec}^{-1})
\]

Substitution in eq 16 leads to

\[
D' = 11.48 \log_{10}(V_{10}) + 0.224 V_{10} - 33. \tag{18}
\]

Table III shows the change in \(D'\) with \(V_{10}\) and also the quantity \(e^{kD'}\) which is the ratio of the saltation layer height to the roughness length. For all but weak winds this is several times larger than the usually claimed ratio of the length of roughness elements to \(z_0\) (20 to 30).

The height \(h\) at which the wind velocity should become independent of changes in \(u^*\) (i.e. where \(\partial V_z/\partial u^* = 0\)) is found by means of eq 15 and 17 as

\[
h = 7.24 \times 10^{-4} e^{(5-D')/2.5} V_{10}^2. \tag{19}
\]
DEPOSITION AND EROSION OF SNOW BY THE WIND

Table III. Values of \( D' \) (eq 15), \( e^{kD'} \), and focal height \( h \) (eq 19), as functions of 10-m wind speed.

<table>
<thead>
<tr>
<th>( V_{10} ) (m sec(^{-1}))</th>
<th>( D' )</th>
<th>( e^{kD'} )</th>
<th>( h ) (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>8.28</td>
<td>27</td>
<td>0.050</td>
</tr>
<tr>
<td>10</td>
<td>10.66</td>
<td>52</td>
<td>0.075</td>
</tr>
<tr>
<td>15</td>
<td>11.44</td>
<td>97</td>
<td>0.124</td>
</tr>
<tr>
<td>20</td>
<td>11.82</td>
<td>113</td>
<td>0.189</td>
</tr>
</tbody>
</table>

The values of \( h \) are also listed in Table III; they are roughly one third of those found by Owen (1964) for sand and agree well with the crossover levels observed in the Byrd Station wind profiles for different 10-m wind velocities (Budd, Dingle and Radok, 1966, Fig. 15).

According to Table III the Byrd Station wind data for occasions of drifting snow lend support to Owen's (1964) contention that the saltation layer affects the free air flow like a solid roughness; they show moreover the dependence of \( D' \) on \( u^* \) which Owen suspected from the sand observations. This dependence will be interpreted in the next section as a decrease in the effective roughness of the saltation layer as the state of full suspension is approached with increasing wind speed.

A further and more direct test of Owen's saltation model can be made by means of the measured snow concentration data themselves. It follows from the solid-roughness equivalence of the saltation layer, and from the self-maintaining thickness of that layer for a given shear velocity, that the roughness length must be a function of the wind speed only and must be independent of the snow concentration in the free air stream. Thus in principle the observed behavior of the roughness length for varying snow concentrations provides a combined test of both the basic hypotheses of the Owen model.

The behavior of the roughness length was examined by Budd, Dingle and Radok (1966) to check Liljequist's (1957) contention that an increase in \( z_0 \) with wind speed found at Maudheim can be explained by the presence of drift snow. This clearly is a tricky thing to prove since an increase in wind speed can also produce genuine roughness changes in the ice shelf surface. However, the Byrd Station observations included a substantial number of cases of snow drift over a hard-packed smooth surface, and their separate analysis showed that \( z_0 \) was independent of the snow concentration in the air near the ice (at the 12-cm level) when the wind speed was eliminated as a variable.

Apart from confirming in a rather subtle manner Owen's concept of a self-maintaining saltation layer with thickness depending only on the shear stress, the foregoing supports the view that the raising and lowering of the snow surface is largely independent of the saltation process and reflects rather the snow content of the free air stream and the horizontal mass flux divergence. Suitably selected data from the Byrd Station material for periods of net deposition and net erosion showed their main difference to be in the mean fall velocity, rather than in the total drift snow content. From the transport discussion it is clear that this implies different mass flux rates for the same wind velocity during deposition and erosion.

The saltation mass flux and its dependence on \( u^3 \), the main consequence of Owen's (1964) self-maintaining layer thickness hypothesis, cannot be tested by means of the Byrd Station drift observations since these did not adequately cover the surface drift (which has little significance for the mass budget). However, another consequence of the same hypothesis is that the particle concentration in the saltation layer should be of the order of magnitude of the fluid density. This is in striking agreement with the deductions made earlier in this section from the wind-dependence of the snow concentrations at different levels. The importance of this agreement cannot be overrated. It explains not only the unexpected existence of a snow drift concentration — wind speed
relation (basically not contained in a “relative” relation of the type of eq 1) but also the way in which net accumulation or ablation can take place in snow drift – by a net vertical mass transport through a self-maintaining agitated saltation layer on top of the rising or descending snow surface.

On this argument the factor controlling deposition and erosion of the snow surface must be sought in changes of the mass flux of the drift snow, most of which takes place in the free air above the saltation layer. We therefore have to consider next the relation between the drift regimes of these two regions.

**SALTATION AND SUSPENSION**

Owen’s (1964) model of saltation implies that the transition from saltation to suspension is governed by the condition \( y = r/g \rho \xi = 1 \). For Bagnold’s desert sands this corresponded to wind velocities between 30 and 40 m sec\(^{-1}\), but for snow the suggested limit was found to be as low as 22 m sec\(^{-1}\). This is a velocity not uncommon in blizzards at Byrd Station and distinctly on the low side for some Antarctic stations; in fact it is only a little higher than the annual mean at Port Martin and substantially below the largest monthly mean velocity recorded there (18 and 28 m sec\(^{-1}\), respectively). Measurable drift snow concentrations are found in the lowest few meters for winds of the order of 10 m sec\(^{-1}\), and even for sand there is evidence of suspension at velocities far below Owen’s threshold value. While Bagnold (1941) made no mention of this (his strongest winds reached 18 m sec\(^{-1}\)) the writer can testify to the existence of noticeable sand quantities at eye level in winds of 15 to 20 m sec\(^{-1}\) during gliding operations on the sand dunes of the Kurische Nehrung, a narrow peninsula on the Baltic coast. For this region Exner (1931) has reported measurements of drift sand concentrations, obtained with air-permeable bags at heights up to 50 cm, which he summarized by the equation

\[
s_z = (A e^{mV - B}) e^{-Cz}
\]

For \( s_z \) in g m\(^{-3}\), \( V \) in m sec\(^{-1}\), and \( z \) in m the constants had the values \( A = 6.8; B = 22.5; C = 6.8; m = 0.18 \). The wind velocity \( V \) was not fully specified by Exner but presumably represented that given by a hand-held anemometer for the 1.5-m level, approximately.

Two of the curves defined by eq 20 are shown as broken lines in Figure 2. They have been extended to the level \( h \) at which the slope of eq 20, viz. \( \partial \log s_z/\partial \log z = -11 \), reached the value \( w^* = w/ku^* \) (cf. eq 2) computed from the roughness lengths and particle fall velocity (\( w = 1.5 \) m sec\(^{-1}\) for \( \xi = 0.32 \) mm) reported by Bagnold (1941) for desert sand and similar wind velocities.

Figure 2 is very suggestive of a gradual transition from the saltation layer to one with suspension in which the concentration is governed by eq 2, but sand concentrations at higher levels would be needed to put this beyond doubt. As a working hypothesis it will be assumed that the transition from saltation to suspension in both the space and wind velocity domains is not abrupt but can be described in probability terms. Thus it might be postulated that suspension becomes “established” once the vertical air velocities exceed the particle fall velocities with sufficient frequency to move a substantial number of particles upward against gravity at any one time. Evidently the meaning of “substantial” remains to be defined in this context. While this could perhaps be done in terms of visibility, following Liljequist (1957), Budd, Dingle and Radok (1966), and Mellor (1966), a more direct definition can be linked to the rate at which the proportion of upcurrents in excess of the fall velocity increases with wind velocity. The onset of suspension would then be placed in the velocity range where this rate has its maximum.

This line of thought resembles the approach used by Businger (1965) but avoids his arbitrary assumption of constant and differing snow concentrations in rising and descending currents. The procedure here proposed can be formalized as follows:

\[ \text{\tiny †} \text{Businger also used an unrealistic value for the snow particle fall velocity.} \]
DEPOSITION AND EROSION OF SNOW BY THE WIND

The vertical velocity \( W \) distribution in the atmospheric boundary layer up to heights of the order of 100 m has been shown by Panofski and McCormick (1960) to be adequately represented by a normal frequency distribution with variance \( W^2 \) proportional to \( u^*^2 \) so that formally

\[
\text{Prob } W = N(0, au^*) \quad \text{(21)}
\]

where \( a = 1.1 \), approximately. This distribution can be reduced to the standard normal deviate \( N(0, 1) \) of statistical tables by changing to the non-dimensional variable \( W/au^* \), and the desired proportion of upcurrents exceeding the particle fall velocity is then defined by

\[
2 \cdot \text{Prob} \left[ \frac{W}{au^*}, \frac{w}{au^*} = \left( \frac{k}{a} \right) w^* = 0.36w^* \right] =
\]

\[
= \left( \frac{2}{\pi} \right)^{1/2} \int_{-\infty}^{\infty} e^{-x^2/2} \, dx.
\]

Equation 22 has been used to describe three different particle drift regimes, viz. those of snow, sand and silt. For drifting snow at Byrd Station the average fall velocity near the surface is 0.3 m sec\(^{-1}\) and is sensibly independent of wind speed; also from eq 17 it follows that \( 0.36w^* = 7/V_{10} \) in this case. For Bagnold's sand the shear velocity could be put into the form \( u^* = 0.247 \cdot 0.06 V_{10} \), and with \( w = 1.5 \text{ m sec}^{-1} \) we obtain \( 0.36w^* = 1.5/(0.066 V_{10} - 0.27) \) for sand. Finally silt particles have a mean diameter of 0.028 mm (Flint, 1957) and fall velocities of the order of 8 cm sec\(^{-1}\). Assuming an overall roughness length of 1 cm for the barren terrain generally surrounding silt source regions we obtain \( V_{10} = 17.25u^* \) and \( 0.36w^* = 1.25/V_{10} \) for silt.

Figure 3 shows the probability 22 as a function of wind speed for snow, sand, and silt. The last of these cases is most clear-cut: for a velocity increase of only 4 m sec\(^{-1}\) the proportion of effective suspension currents (exceeding the particle fall velocity) increases from zero to 70%.
For sand on the other hand the increase is quite gradual and winds of 37 m sec\(^{-1}\) are needed to raise the proportion of suspension currents to 50%. For snow the percentage of such currents rises sharply for winds between 4 and 12 m sec\(^{-1}\) and passes the 40% level for \(V_{10} = 9\) m sec\(^{-1}\) where common experience shows snow drift to start extending to higher levels. This point might then be used as an arbitrary definition of the effective onset of suspension in snow drift. The main point of Figure 3 is, however, to demonstrate that snow appears to have its transition from saltation to suspension drift at the wind velocities most common near the earth's surface, whereas for silt only the suspension regime and for sand only that of saltation are normally encountered.†

For snow the physical relevance of the probability argument can be demonstrated more fully by means of the mean particle fall velocities established for different levels and wind speeds from the Byrd Station measurements (Budd, 1966). Figure 4a shows these fall velocities in the form given by Budd while in Figure 4b they have been replotted on a scale of

\[
(2/\pi)^{1/2} \int_{-\infty}^{\infty} e^{-x^2/2} \, dx.
\]

\[
\bar{w}_z/1.1\bar{u}^* = 0.36\bar{w}^*
\]

† Special sands such as those of the White Sands National Monument in New Mexico remain to be studied along these lines.
DEPOSITION AND EROSION OF SNOW BY THE WIND

Figure 4. Mean fall velocities for Byrd Station snow drift particles (a, after Budd, 1966), and the corresponding suspension up-current frequencies (b), for two wind velocities.

This brings out the fact that the mean fall velocities correspond to suspension current percentages of the order of 60 to 70% irrespective of height and wind speed changes. The onset of suspension is however marked as taking place at lower percentages by the way the observed fall velocities in Figure 4b rotate with increasing wind speed.

A final supporting point is provided by the decrease in the effective roughness of the saltation layer with increasing wind speed, which was noted in the previous section (Table III). This can now be interpreted as reflecting the increasing number of particles which move into and out of the saltation layer, rendering its upper boundary more and more diffuse. This has some similarity to what happens at the surface of a liquid approaching its boiling point, and the transition from the saltation to the suspension regime might thus be regarded and treated in kinetic terms as a kind of phase change. Some implications of this for the study of particle transport in tubes will be taken up in the last section. To complete the discussion of unconfined snow drift the factors responsible for changes in the horizontal mass flux of suspended snow must now be considered.

OBSTACLES AND SNOW DRIFT

Snow deposition and erosion near an obstacle is often blamed intuitively on eddies created by the obstacle. Closer reasoning (Odar, 1965) has been concerned above all with the surface shear stress as a crucial factor governing deposition or erosion. Neither line of thought seems capable of accounting for the actual amount of snow involved which on the argument and evidence presented here should be explainable solely in terms of minor variations in the horizontal drift flux past the obstacle.
Recent theoretical work on boundary layer flow around obstacles (Stratford, 1959a, b; Townsend, 1961, 1965a, b, c) has established the existence of two major flow regions in such cases. In the first of these the flow structure has not yet been affected by the obstacle or equivalent change in surface conditions, and its properties depend on those of the entire flow, including its most distant portions (this is a general characteristic of free turbulence). The second region is that close to the obstacle where the flow is completely modified and dominated by local conditions. A schematic view of this dichotomy is shown in Figure 5 which has been reproduced from Townsend (1965c) and refers to the case of an adverse pressure gradient, such as would be encountered by the flow some distance ahead of an obstacle.

It is instructive to compare this theoretical picture with one obtained from extensive measurements by Plate and Lin (1965) of the flow across two-dimensional obstacles of simple shape in the substantial boundary layer that can be achieved in the Micrometeorological Wind Tunnel Facility of the Colorado State University (for a description see Plate and Cermak, 1963). Figure 6 has been reproduced from Plate and Lin’s paper and brings out clearly the complex structure of the region intermediate to the essentially undisturbed and fully modified layers which are amenable to theoretical analysis. Flow measurements are clearly needed in the intermediate region not only behind but also upstream of a variety of obstacles as a first prerequisite for the understanding of their effects on drifting snow.

A combination of wind tunnel and field studies promises to be especially useful in this context. Thus wind tunnel observations of the flow around snow fences (Kreutz and Walter, 1956) can be matched with measurements of the full scale wind field produced by a hedge with bottom gap (Rider, 1952). In a wind tunnel study of the flow around the buildings for the new Wilkes Station of the Australian National Antarctic Research Expeditions (Melbourne, 1966) the measurements were obtained outside the boundary layer, and their full relevance for the case of natural flow has
yet to be confirmed by observations of the actual deposition and erosion patterns. Many such observations are available for various elevated structures (e.g., Kovacs, unpublished) but cannot be interpreted with any degree of assurance in the absence of adequate wind data. There remains an urgent need for a comprehensive program of coordinated wind tunnel and field flow measurements relating to obstacles affected by snow drift or used in their creation and prevention, and some suggestions for such a program will be made at the end of this paper.

Pending work along such lines the hypothesis linking the deposition and erosion of snow by the wind primarily to variations in the horizontal mass flux can be tested by a fairly crude calculation for the snow drifts which form regularly near the DEWline radar stations on the Greenland Ice Cap (Mellor, 1965). These snow accumulations have not yet been studied in the course of formation and normally are not left to grow unchecked. However there are reports that they can build up to something like their usual end-of-season proportions in the course of a single three-day storm. This makes it possible to compare the observed volume of deposited snow with the horizontal drift transport and its likely divergence upwind of the radar building.

For the calculation of the expected snow deposit size will be made of Rider's (1952) wind profiles upwind of a 1.68-m-high hedge with open structure in the lowest 25 cm. These winds are shown in non-dimensional form in Figure 7a for three levels (measured like the horizontal distance in obstacle height units) and have been shown by Townsend (1965b) to be in good agreement with those expected from boundary layer theory. They refer to a scale very different from that of the radar buildings (which are generally at least 20 m high, 40 m wide, and elevated some 3–5 m above the snow surface) but a comparison with flow velocities obtained in the wind tunnel experiments of Plate and Lin (1965) and also plotted in Figure 7a suggests that the scale difference is much less critical than the presence or absence of the bottom gap.

The slowing down of the air in line with and above the obstacle may be taken as represented roughly by the broken straight line in Figure 7a which has the equation

$$\log \left( \frac{V}{V_\infty} \right) = 0.1 \log \left( \frac{x}{l} \right) \quad (23)$$

where $V_\infty$ is the unaltered free stream velocity at the level in question and $x$ is the horizontal distance from the obstacle. The horizontal velocity gradient upwind of the obstacle is then given by

$$\frac{\partial V}{\partial x} = \frac{1}{\ln \left( \frac{x}{l} \right)} \frac{1}{0.1} \quad (24)$$

The latest Russian information on the subject given by Dinnin (1963) in a recent book on snowstorm mechanics has not yet received adequate study at the time of writing but although very substantial and relevant does not seem likely to remove the need for such a program.
The effect of the retardation described by eq 24 will be to create an elevated drift snow "source" upwind of the obstacle. The snow deposition results through turbulent diffusion from this source, but the complexity of the (essentially three-dimensional) flow field and the lack of information about the transfer processes in the disturbed flow rule out the use of formal results such as those given by Rounds (1955). Instead a greatly simplified procedure is suggested by the facts that 1) a few of the snow particles at higher levels can reach the surface during the interval in which the air passes the obstacle, and 2) the bulk of the drift snow is transported in the lower layers in which, moreover, a downward velocity component created by the obstacle will assist the
deposition. Leaving aside the important but intractable complication that the source layer will increase in vertical extent as the obstacle is approached, the deposition rate $S$ follows from the continuity equation for the drift content of the layer

$$\frac{\partial \Gamma}{\partial t} + \frac{\partial (\Gamma V)}{\partial x} - S = 0 \quad (25)$$

where $\Gamma$ and $V$ are related by $\Gamma V = Q = \int n V \, dz$, the integral being extended from the height of the bottom gap to somewhat above that of the center line through the obstacle.

For steady-state conditions $\partial \Gamma/\partial t = 0$ and limits for $S$ can be established as follows. If the average drift density does not change appreciably during the flow of the air past the obstacle

$$S_1 = \Gamma_\infty \frac{\partial V}{\partial x} = \frac{Q_\infty}{V_\infty} \frac{\partial V}{\partial x} = \frac{Q_\infty}{10^{1.1}x^{0.9}} \quad (26)$$

At the other extreme we have the case where the horizontal drift transport has become adjusted everywhere to the local wind velocity in accordance with the empirical transport-wind relationships of the form $\log Q = a + bV$ established earlier. Then

$$S_2 = \frac{\partial Q}{\partial x} = \frac{\partial Q}{\partial V} \frac{\partial V}{\partial x} = Q \frac{\partial \log Q}{\partial x} \frac{\partial V}{\partial x} = 2.3b V_\infty \frac{10^b(V-V_\infty)}{10^{1.1}x^{0.9}} \quad (27)$$

where $b$ is a constant of the order of 0.1 and the factor 2.3 represents $\log_e 10$.

Relations 26 and 27 give an idea of the relative magnitudes of horizontal drift transport and deposition in the upwind retardation zone where $x$ has values between 1 and 10. Moreover, since $V_\infty$ can reach 40 m sec$^{-1}$ or more the estimate (eq 27) based on a completely adjusted horizontal drift transport can be 4 times that assuming no adjustments to the drift content in the retardation zone.

For conservative estimates of the snow accumulation to be expected upwind of the radar buildings the quantity $S_1$ was computed for two representative wind speeds (10 m sec$^{-1}$ and 20 m sec$^{-1}$) using for $Q_\infty$ one tenth of the respective transport rates established from the Byrd Station measurements for the layer between the surface and 300 m. Partial transports such as shown in Figure 1 would evidently be more relevant in this connection, but without factual information on the winds and drift snow concentrations in question no such refinement is called for.

The deposition profiles computed from eq 26 for three-day storms are shown in Figure 7b; to convert the computed water equivalents to deposited snow heights they have simply been interpreted as inches of snow instead of cm of water. The heights of the snow drift upwind of the radar station Dye 3 for the end of the first winter (before snow removal had changed the initial conditions appreciably) have been scaled from Figure 62 of Mellor's (1965) survey and are given for comparison with the computed profiles in Figure 7b.

The computed values for 20 m sec$^{-1}$ are close to the observed profile, but in view of the rough argument used this must be regarded as somewhat fortuitous. Nevertheless the right order of magnitude has been achieved by the calculation and with it support for the view that the free air flow field associated with an obstacle must be given the principal attention in any study aimed at explaining the deposition and erosion pattern associated with the obstacle.
CONCLUSIONS AND SUGGESTIONS

The result and conclusion of the preceding section has a direct bearing on the planning of wind tunnel tests which have been proposed in connection with the Dye site snow drifts. The Micrometeorological Wind Tunnel Facility at Fort Collins in which these tests will be carried out offers the possibility not merely of exploring the entire flow fields both upstream and downstream of the model radar buildings but also of varying the hydrostatic stability of the air current. This stability may well play an important role in determining the shape and limiting size of the real drift. It is well known (Exner, 1931) that the crescent shape of sand dunes arises from a combination of minimum forward movement of the central section of maximum thickness, and deflection of sand grains down the sides of the dune. For snow dunes that deflection would be accentuated by the presence of intense surface inversions such as are commonly observed on the ice cap even during periods of strong winds, and snow deflection could then produce in effect a terminal dune size beyond which the growth rate is inappreciable.

For the free air stream measurements and their interpretation the possibility of a zero surface stress profile arising from natural causes or through judicious shaping of the upstream topography should be kept in view. The work of Stratford (1959a, b) and Townsend (1961, 1962, 1965 a—c) has established that this case represents the minimum interference of the obstacle with the free air flow, clearly the basic objective of any attempt at controlling snow deposition† if, as has been argued in this report, such deposition results primarily from a convergence of the suspension drift flux.

In view of the restricted validity of even the most sophisticated wind tunnel experiments for the equivalent processes in the atmosphere parallel field experiments are highly desirable and should form an intrinsic part of the snow drift program now under consideration. Pilot experiments with medium-scale field models have already been carried out at Lebanon, N.H., Regional Airport (flow profiles) and in Greenland (accumulation measurements) by Tobiasson and Reed (unpublished) but these must be combined to yield the full value of this approach. Systematic measurements on the actual Dye site drifts would also be highly desirable; even simple profiling by theodolite or Abney level a few times during their formation would provide, together with wind records, invaluable material for checking the wind tunnel results and guiding their interpretation.

Saltation and the surface wind stress which keeps the snow in a state of agitation form integral parts of the drift process and require further study. In the case of saltation this calls for field work which could at the same time serve as one starting point of an investigation of the electrical phenomena associated with blowing snow. These have been shown both by field measurements (Wishart and Radok, 1967) and in the laboratory (see especially Latham and Stow, 1967) to arise largely if not wholly from differences between the temperature of the snow surface and that of the drift particles. There is a distinct possibility that a better understanding of the electrical phenomena may clarify processes in the saltation layer in general which are difficult to study directly.

Although saltation probably does not play the major role in snow deposition and erosion by the wind it may well hold the key to problems of pneumatic particle transport in tubes and channels. Theoretical work in this field is needed initially in connection with Bagnold’s more recent semi-empirical studies (Bagnold, 1954, 1955, 1956, 1961); the work of Owen (1964) which has been extensively referred to in this survey points the way. In addition to such theoretical work the discussion on saltation and suspension in this paper suggests a direct experimental approach with measurements of transport rates, for particles in the drift or chip size ranges and in different concentrations, moving through tubes of a given cross-section area but differently proportioned rectangular cross sections. In this way the relative importance of the saltation and suspension modes could be varied systematically to study their effect on the transport rate.

As an extension of these measurements (arising from the phase analogy pointed out at the end of the section on Saltation and suspension) the inclination of the tubes upward or downward would be

† This is essentially identical with a suggestion by Mellor (1965, footnote p. 52).
equivalent to a variation in the restoring force at the top of the saltation layer; alternatively particles of different density could be used to study the same effect. Such measurements will be useful in themselves but also serve to test any transport theory indicating the minimum stress needed to keep a given particle concentration on the move.

**LITERATURE CITED**


DEPOSITION AND EROSION OF SNOW BY THE WIND


The theories of uniform and non-uniform drifting snow are summarized with special emphasis on drift transport as a function of wind velocity. Using the work of Owen (1964) and the observations of the Byrd Station Snow Drift Project (Budd, Dingle and Radok, 1966) it is confirmed that the snow drift process involves a mobile surface layer of saltating particles, with a self-regulating thickness depending only on the surface stress and not on the snow concentration in the free air stream. It is shown to be a characteristic of snow (in contrast to sand or silt) that saltation and suspension drift occur side by side and that the latter reaches predominance as the wind velocity rises through the most common range of surface values. Theoretical reasons and observational evidence are produced for the view that deposition or erosion occurs on the snow surface during snow drift primarily as the result of mass flux convergence or divergence in the free air stream. This implies that the associated vertical mass flux penetrates the saltation layer which moves up or down with the snow surface. The survey concludes with suggestions for the experimental study of snow deposition and erosion in terms of the free air flow field and for a study of pneumatic particle transport in terms of saltation and of its electrical effects.
<table>
<thead>
<tr>
<th>KEY WORDS</th>
<th>LINK A</th>
<th>LINK B</th>
<th>LINK C</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ROLE</td>
<td>WT</td>
<td>ROLE</td>
</tr>
<tr>
<td>Drifting snow</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Snowdrift</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Saltation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Snow surface</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Snow deposition</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Snow erosion</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Blowing snow</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>