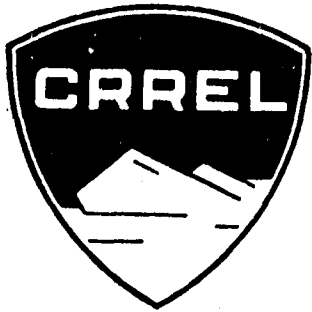


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Cold Regions Science and Engineering Part III, Section A3c

BLOWING SNOW

by
Malcolm Mellor

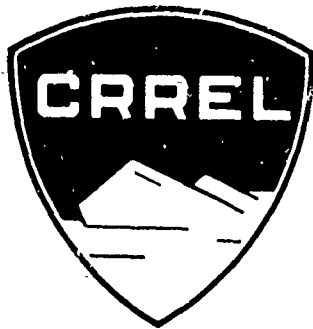
NOVEMBER 1965

U.S. ARMY MATERIEL COMMAND
COLD REGIONS RESEARCH & ENGINEERING LABORATORY
HANOVER, NEW HAMPSHIRE



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Part III, Section A3c**

BLOWING SNOW

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NOVEMBER 1965

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DA Project IVO25001A130



PREFACE

This monograph summarizes available information on blowing snow.

USA CRREL is an Army Materiel Command laboratory.

DA Project IV025001A130

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EDITOR'S FOREWORD

"Cold Regions Science and Engineering" consists of a series of monographs written by specialists to summarize existing knowledge and provide selected references on the Cold Regions, defined here as those areas of the earth where the freezing and thawing of soils is an essential consideration in engineering.

Sections of the work are being published as they become ready, not necessarily in numerical order but fitting into this plan, which may be amended as the work proceeds:

I. Environment

A. General

1. Geology and physiography
2. Permafrost - perennially frozen ground
3. Climatology
 - a. Cold Regions climatology I
 - b. Cold Regions climatology II
 - c. Antarctic climatology
4. Vegetation

B. Regional

1. The Antarctic ice sheet
2. The Greenland ice sheet

II. Physical Science

A. Geophysics

1. Heat exchange at the earth's surface
2. Exploration geophysics

B. Physics and mechanics of snow as a material

C. Physics and mechanics of ice

1. Snow and ice on the earth's surface
2. Ice as a material

D. Physics and mechanics of frozen ground

III. Engineering

A. Snow engineering

1. Engineering properties of snow
2. Construction
3. Technology
 - a. Explosions and snow
 - b. Snow removal and ice control
 - c. Blowing snow
 - d. Avalanches

4. Oversnow transport

B. Ice engineering

C. Frozen ground engineering

1. Site exploration and excavation
2. Buildings, utilities and dams
3. Roads, railroads and airfields
4. Foundations
5. Sanitary engineering

D. General (construction, etc. not covered by III C)

BLOWING SNOW

by

Malcolm Mellor

Introduction

Wind-blown snow causes a variety of problems over the great areas of the earth where it is commonly experienced. While snow is actually blowing, outdoor activities, including transport, are curtailed and electrical disturbances, such as radio noise, are experienced. At the end of a period of blowing snow, many areas (e.g., roads and avalanche paths) are found to be covered by unwanted deposits of snow, while other places (such as arable land or ski trails) may have been denuded of much-wanted snow.* Redistribution of snow by winds is also a factor to be considered in the study of glaciers and the management of drainage basins.

Simple measures for the control of blowing snow, notably snow fences, have long been used, but the methods currently available for abating the nuisance of blowing snow (or even turning it to advantage) are far from adequate in a period of increasingly vigorous human activity in snow-covered regions. Fortunately there are in progress numerous studies which should lead to an early improvement in the situation. Field studies are yielding data which clarify the physical situation, while the general theory of two-phase flow (i.e., fluids and solid particles) is exciting interest in a number of technical fields. Finally, techniques for the execution and interpretation of model and prototype investigations on drift control measures are being refined.

Sources of snow particles

Snow particles carried by the wind may be supplied directly by precipitation falling during the wind storm, or they may be particles deposited on the surface earlier and subsequently picked up by wind. In considering the transport of particles by turbulent diffusion it does not matter greatly whether the particles fall into the wind stream from above or whether they are picked up from the surface, but the distinction must be made in view of the widespread adoption of the "threshold velocity" concept (the commonly held view that snow is picked up from the surface and blown along when the wind speed exceeds a definite critical value).

In all wind conditions of practical interest, including even light breezes, flow near the ground is turbulent, and small particles (or slow-falling dendritic crystals) can be carried along with a mean path which is close to horizontal. If precipitation is in progress, then snow can be transported in a horizontal direction even at low wind speeds. If, on the other hand, no new snow is falling, appreciable wind speeds may be necessary to apply sufficient shear and lift forces to dislodge particles from the surface and diffuse them into the air stream. The actual wind speed needed to disturb the surface and transport particles depends on the condition of the surface snow: if the snow is loose and unbonded, fairly low wind speeds (say 3 to 8 m/sec at 10-m height) will suffice, but if the surface snow is densely packed and firmly bonded by thaw-freeze or sintering (age-hardening), it may resist very strong winds (more than 30 m/sec). Figure 1 gives Russian field data relating snow "hardness"† to the wind speed necessary for dislodging that snow. Actually the disruption of a snow surface should properly be related to the surface shear stress, which is dependent on surface roughness as well as wind speed, but it will be seen from the data presented later that the distinction is not of great practical importance.

*At Schefferville, Canada, snow drifting is being induced in an effort to raise ground temperatures as a prelude to mining.

†A strength index obtained from measurements with a spring-loaded penetrometer.

Reported "threshold velocities," above which snow transport is said to begin, can be misleading if they are not referred to the source of supply for snow particles. Perhaps more consistent results would be obtained if threshold velocity were to be defined as the wind speed below which snow transport ceases. The availability of loose snow particles in a particular size range will also determine whether or not a wind stream is "saturated," i. e., whether it is actually carrying all the snow it is capable of carrying.

Surface winds

As wind blows across a snowfield, drag against the surface creates turbulence and produces a velocity gradient in a direction normal to the surface, i. e., there is a turbulent boundary layer. Turbulence develops at low wind speeds—a few meters per second—but generally the flow does not become fully turbulent until the velocity at standard anemometer height of 10 m exceeds about 10 m/sec⁵³.

With neutral stability (negative temperature gradient of approximately 1C/100 m) it has been found that there is a logarithmic relationship between wind speed and height above the surface, i. e., logarithm of height is directly proportional to velocity at that height. If temperature gradients differ significantly from the adiabatic gradient of 1C/100 m, the stratification of the air becomes stable or unstable; a larger negative gradient ("super-adiabatic lapse rate") leads to instability, while a smaller negative gradient or a positive gradient ("inversion") gives stability. Under these conditions the logarithmic velocity profile no longer represents the conditions adequately, since a plot of log. height (ordinate) against velocity (abscissa) is a curve—convex up for stable equilibrium and concave up for unstable conditions.*

Field measurements show that with typical blowing snow conditions the logarithmic wind profile represents the measured velocity distributions satisfactorily^{53, 19, 12} and therefore it may be adopted for snowdrifting studies:

$$u = \frac{u_*}{k} \ln\left(\frac{z}{z_0}\right) \quad (1)$$

where u is mean wind speed at height z above the surface, z_0 is a surface roughness parameter, k is von Kármán's constant (≈ 0.4 for snow-free air but possibly smaller when solid particles are present); u_* is a reference velocity termed the "friction velocity," and defined in terms of the surface shear stress τ (see below) and the air density ρ by

*In these circumstances Deacon¹⁸ proposes the relation

$$u = \frac{1}{1-\beta} \frac{u_*}{k} \left[\left(\frac{z}{z_0}\right)^{1-\beta} - 1 \right]$$

in which $\beta > 1$ for superadiabatic gradients
 $\beta = 1$ for adiabatic gradients
 $\beta < 1$ for inversions.

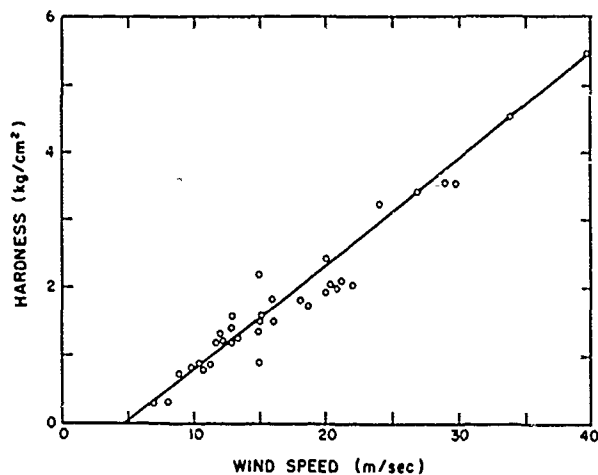


Figure 1. Relationship between snow "hardness" and the wind speed necessary for erosion of that snow. (After Kotlyakov⁴⁸)

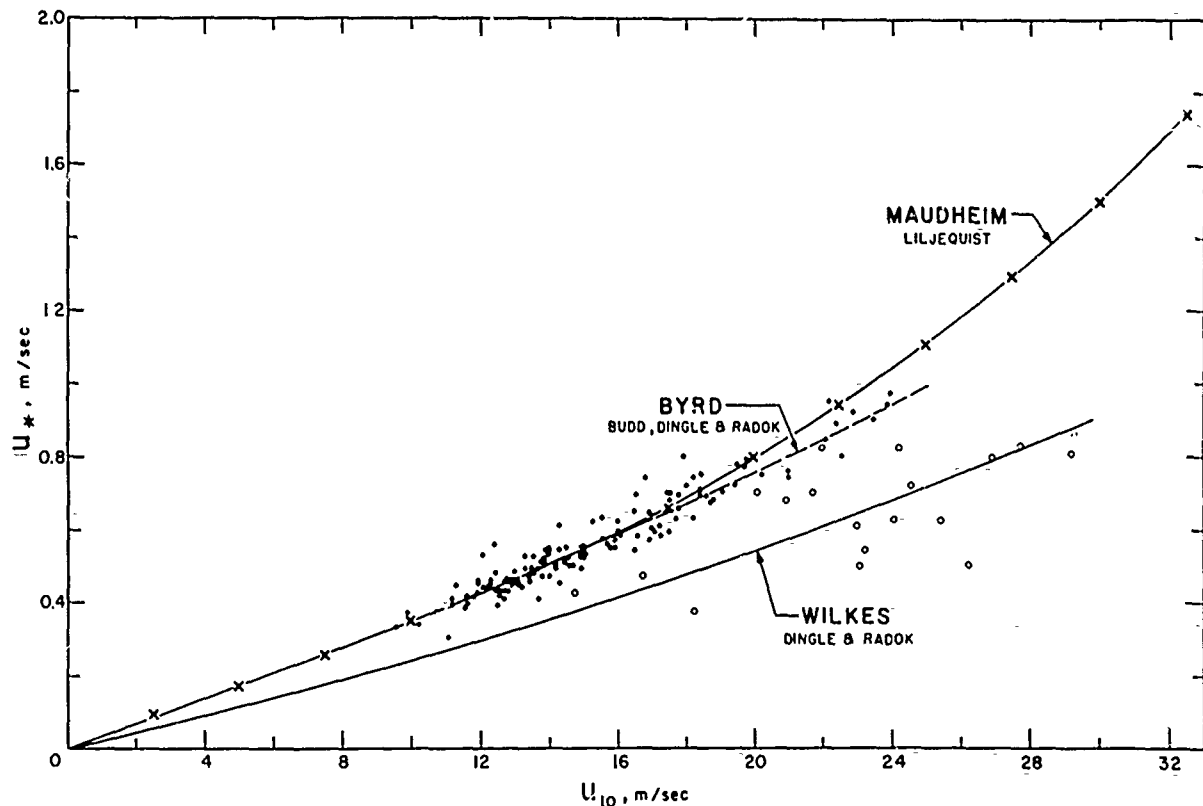


Figure 2. Friction velocity u_* as a function of wind speed at 10-m height u_{10} . (Data from ref. 12, 19, 53).

$$u_* = \sqrt{\frac{\tau}{\rho}} \quad (2)$$

As can be seen from eq 1, u_* measures the slope of the logarithmic wind profile. Its value depends on the roughness length z_0 , which itself is given by the intercept of the logarithmic wind profile.

Values of u_* measured over snowfields are given as a function of wind speed at anemometer height u_{10} in Figure 2; u_* is about 3 to 4% of u_{10} .

The roughness length z_0 for snow surfaces is relatively small; the most frequently reported values are of the order of 10^{-2} cm, but values as high as 10^{-1} cm and as low as 10^{-6} cm have been reported. A significant fact is that the roughness length for snowfields is apparently unrelated to the height of gross surface features such as dunes and sastrugi¹²; it seems to decrease slightly with the "hardness" of the surface. There are reasons to believe that z_0 should increase slightly with wind speed, but the available data show little evidence of this. It has been suggested by Liljequist⁵³ that z_0 might vary with the concentration of suspended snow, but Budd, Dingle and Radok¹² have presented an argument to show that this is not so. Some values of z_0 are given in Table I.

Taking Schlichting's criteria for smooth and rough flow (Sutton⁸⁸) and inserting values for u_* and z_0 from Figure 2 and Table I, it is found that in almost all cases for strong winds, snow surfaces are aerodynamically rough (no laminar sub-layer); some small roughness lengths measured by Dingle and Radok (but apparently of doubtful reliability) indicate possible exceptions.

Table I. Surface roughness parameter for snowfields.

Liljequist ⁵³	2.5 to 13 m/sec, no blowing snow	10^{-2}
Liljequist ⁵³	13 to 32.5 m/sec, blowing snow	1.5×10^{-2} to 5.8×10^{-1}
Priestley ⁵³		10^{-1}
Dingle and Radok ¹⁹	14.7 to 29.2 m/sec, blowing snow	10^{-6} to 2×10^{-1}
Sverdrup ⁸⁹		2.3×10^{-1}
Budd, Dingle and Radok ¹²	10 to 24 m/sec, blowing snow: Average for hard smooth snow Average for dunes, barchans and sastrugi Average for soft snow	1.19×10^{-2} 1.46×10^{-2} 2.20×10^{-2}
Dalrymple <u>et al.</u> ¹²	5 to 10 m/sec	10^{-3} to 10^{-2}
Rusin ⁸⁰	0 to 20 m/sec	7×10^{-3} to 3.7×10^{-2}

The wind shear stress τ is generally assumed constant to a height of at least 25 m; by analogy with Newtonian viscous shear, τ is related to the vertical gradient of (temporal) mean velocity in the form

$$\tau = K \rho \frac{\partial u}{\partial z}. \quad (3)$$

K is the eddy viscosity, or momentum exchange coefficient for turbulent mixing, which derives from Prandtl's observation that the downward flux of momentum in a turbulent fluid is manifested as shear stress in the surface layers. Prandtl introduced a "mixing length" l (analogous to, but much larger than, the mean free path of gas theory) such that

$$\frac{\tau}{\rho} = u_*^2 = l^2 \left(\frac{\partial u}{\partial z} \right)^2. \quad (4)$$

It is assumed that l is directly proportional to height, k being the proportionality constant:

$$l = kz. \quad (5)$$

Equations 4 and 5 give

$$\frac{\partial u}{\partial z} = \frac{u_*}{kz} \quad (6)$$

the solution of which is eq 1, with z_0 the constant of integration. The eddy viscosity K can be expressed as

$$K = k u_* z. \quad (7)$$

The theory for turbulent suspension of snow particles, discussed later, treats the vertical transport of particles as analogous to momentum flux, and an eddy diffusivity for particle transport corresponding to the eddy viscosity K is introduced.

The thickness of the boundary layer, in which velocity increases from practically zero at the surface to the free stream velocity, is difficult to define. Even in an idealized flow, boundary layer thickness is necessarily defined arbitrarily; one simple definition gives the limit of the boundary layer as the distance from the surface

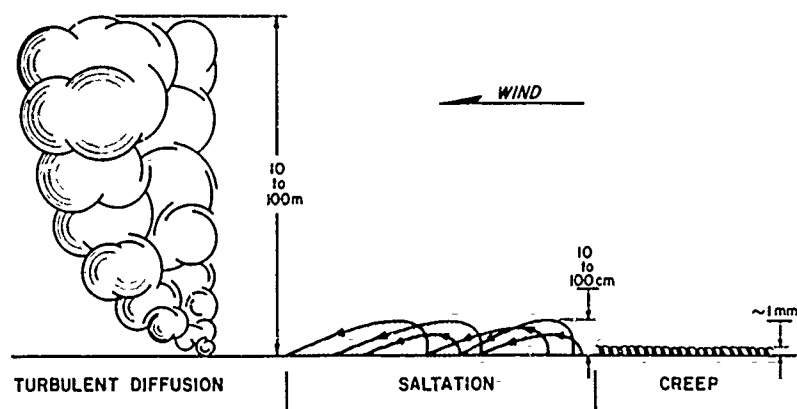


Figure 3. The three modes of transport for blown snow.

at which the velocity defect is 1% (i. e., the velocity is 99% of the free stream velocity). In some meteorological problems boundary layer thickness is taken in excess of 100 m, but for certain practical aspects of the blowing snow problem it may suffice to consider a boundary layer of the order of 10 m thick, although the velocity defect at 10-m height exceeds the 1% level. Lower layer wind profiles for snowfields are given by Liljequist⁵³ and Budd *et al.*¹²

Wind profiles are measured by exposing several anemometers at different levels on a single mast, the vertical intervals between instruments being arranged to give approximately equal spacing on a logarithmic scale. The slope and intercept of a plot of log height against velocity yield, for the logarithmic profile, u_{*} and z_0 . The wind profile may be extended to higher levels by balloon tracking, or by observing the distortion of a near-vertical smoke trace laid by a rocket.¹²

As an alternative to the logarithmic relationship for the wind profile, a power law is still occasionally used for mathematical convenience. In this relation velocity increases with a power of the height, the value of the exponent for conditions of neutral stability being approximately 1/7.

Snow transport mechanisms

In studies of wind-blown sand and other fluid-borne sediments, three transport mechanisms have been distinguished (Fig. 3). They are:

(a) Creep, in which particles migrate along the surface while still in contact with the surface most of the time.

(b) Saltation, in which particles bound along the surface; after traveling a curved trajectory under the influence of wind and gravity forces, each particle is thought to rebound or to eject other particles following impact.

(c) Turbulent diffusion, in which particles are held in suspension by vertical mixing, so that they travel in the air stream without necessarily contacting the ground.

The "creep" process may be observed in dry snow when particles roll along the surface impelled by wind shear. It is not likely to be of great practical significance in drift control but, because of its probable importance in the formation and migration of snow ripples and sastrugi, it is a process which merits investigation. (It appears to have been neglected so far.)

A mode of snow transport identifiable as "saltation" is readily apparent when brisk winds (≈ 6 to 10 m/sec) blow across surfaces of cold, loose snow. Particles

stream along near the surface in a layer about 10 to 20 cm deep, obscuring the smaller surface features* and sometimes the feet of the observer. Saltation has been studied by gauging snow transport rates in the lowest layers, by photographing particle trajectories by streak methods, and by calculating particle trajectories, using graphical and numerical methods based on certain theoretical assumptions. From the limited findings published so far (see p. 7) it is difficult to determine the true nature and the practical significance of saltation. Most observers feel that maximum trajectory heights for appreciable concentrations of saltating particles are in the range 10 to 100 cm, although this may be something of an arbitrary limit, as will be seen from a following argument. In some areas frequent low drifting creates troublesome snow accumulations on highways and in other critical places; this is the type of snow transport which is classed as saltation.

Turbulent diffusion probably becomes the dominant mechanism for snow transport when appreciable concentrations of snow particles are carried along at and above eye level. In major polar snowstorms most of the snow is transported by turbulent diffusion, since the snow-bearing layer may exceed 100 m in vertical thickness. This mechanism is of prime importance in the consideration of snowdrift formation around tall objects, such as engineering structures. It is also the principal mechanism affecting wind distribution of snow on glaciers and in the mountains.

The mechanics of particle motion in a turbulent fluid involves complex problems which cannot yet be solved without making many assumptions. In the following summary of available information, greatly simplified treatments are adopted; for more rigorous discussion of the fundamentals the reader is referred to the literature on two-phase flow.

Mechanics of saltation

It is said that the saltation motion prevailing in low-level sandstorms depends upon elastic impact between particles at the surface³. Each leap of a particle is supposed to be initiated primarily by elastic rebound from the surface or by ejection of the particle following impact from another particle. This explanation of the motion has been extended, somewhat uncritically, to saltation of snow.

Since a mass of loose snow is in most respects inelastic, the elastic collision theory for saltation must be questioned. In a simple desk-top experiment with glass beads of 1 mm and 4 mm diam, maximum heights of bounce from deep beds of glass beads were more than an order of magnitude smaller than bounce heights from a glass plate. This suggests that the energy of the impact was absorbed by internal friction when many particles in the impact zone were slightly displaced. A more significant experiment might be conducted by shooting glass beads at shallow angles ("glancing incidence") to a bed of beads and photographing the resulting trajectories. Until the "bouncing" hypothesis is tested experimentally, it seems desirable to keep in mind the possibility that snow grains agitated by wind shear may be lifted into the air by surface eddies. With these cautionary statements, existing ideas and information on saltation can be summarized.

According to current notions, a saltating particle is ejected, or bounced, from the wind-agitated surface and into the air stream. The speed and direction of the particle on ejection are dependent on the energy and geometry of the previous impact, and on any aerodynamic lift forces which might be developed. Assuming that initial lift forces are not sustained after ejection, the forces acting on the particle during flight are its weight and the aerodynamic drag, which is a function of the particle dimensions and the particle velocity relative to the fluid. The position and vector velocity of a particle at any given instant after take-off are determined by the history of

*Seen from the air, saltating snow appears to lower the reflectance of a snowfield, an effect which should in principle become more marked as the thickness of the blowing snow layer increases. Such an effect might significantly affect the radiation balance.

the motion up to that time. In essence, determination of a saltation path is a complicated ballistics problem with an initial probability consideration.

In obtaining a solution for saltation trajectories, the chief difficulty is presented by the aerodynamic drag, which cannot even be specified conveniently for a spherical particle in a laminar boundary layer, let alone a spinning angular grain entering a turbulent flow. Simplifying assumptions permit attempts at solution to be made, however, using graphical or numerical methods of successive approximation.

Bagnold² described a graphical technique for constructing sand particle trajectories; this method was applied to snow by Mellor and Radok,⁶¹ and was subsequently modified for numerical (computer) solution by Jenssen⁴¹. Strom et al.⁸⁷ made an independent approach and derived equations which can be solved numerically. Wind tunnel tests with glass beads confirmed experimentally a predicted trajectory. In considering bed-load transport by fluids, Yalim⁹² proceeded from a dimensional analysis of the variables to development of the mechanics for average grain motion in saltation, finally arriving at an expression for the mass of solid particles transported. Results of this study have not been applied to snow so far.

Turbulent diffusion

Snow particles in turbulent air are carried along by the wind and diffused by eddies, while at the same time they tend to settle under gravity. Since the particles are small compared with the scale of air turbulence, a simplified discussion of their suspension can be based on the general theory for turbulent diffusion of scalar properties or fine aerosols (see, for example, Hinze³⁸ and Fuchs³⁰).

Consider the elementary section of an air stream shown in Figure 4. The concentration of snow particles in the element is \bar{n} , and horizontal and vertical components of particle velocity (temporal means) are \bar{u} and \bar{v} respectively. At a later stage \bar{u} and \bar{v} will be related to the wind velocity and the fall velocity of a particle through air respectively. Concentration \bar{n} decreases in the positive \bar{x} and \bar{z} directions, so that concentration gradients are negative. Mass transfer coefficients (eddy diffusivities) in \bar{x} and \bar{z} directions are ϵ_x and ϵ_z .

Mass flux into and out of the element through each face is shown in Figure 4, and accumulation in the element is $(\partial n/\partial t)dxdz$. Equating net flux and accumulation:

$$-\frac{\partial}{\partial x} \left(\bar{u}n - \epsilon_x \frac{\partial n}{\partial x} \right) - \frac{\partial}{\partial z} \left(\bar{v}n - \epsilon_z \frac{\partial n}{\partial z} \right) = + \frac{\partial n}{\partial t} \quad (8)$$

Since concentration gradients in the longitudinal direction are usually far smaller than those in the vertical direction, diffusion in the \bar{x} direction can be neglected, reducing eq 8 to

$$-\frac{\partial}{\partial x} (\bar{u}n) - \frac{\partial}{\partial z} \left(\bar{v}n - \epsilon_z \frac{\partial n}{\partial z} \right) = + \frac{\partial n}{\partial t} \quad (9)$$

Written for the plane $z=0$, this gives the rate of deposition or deflation.

For steady-state conditions (concentration at any level invariant with time) the accumulation $\partial n/\partial t$ is zero, and for a situation where wind speed does not change with distance \bar{x} the flux gradient $(\partial/\partial x)(\bar{u}n)$ is also zero. Hence with these stated conditions there is a further simplification of eq 9 to a one-dimensional form which gives the distribution of particle concentration with height:

$$\bar{v}n - \epsilon_z \frac{\partial n}{\partial z} = 0. \quad (10)$$

In solving eq 10 for analysis of field data it has been the practice to make further assumptions, all somewhat questionable:

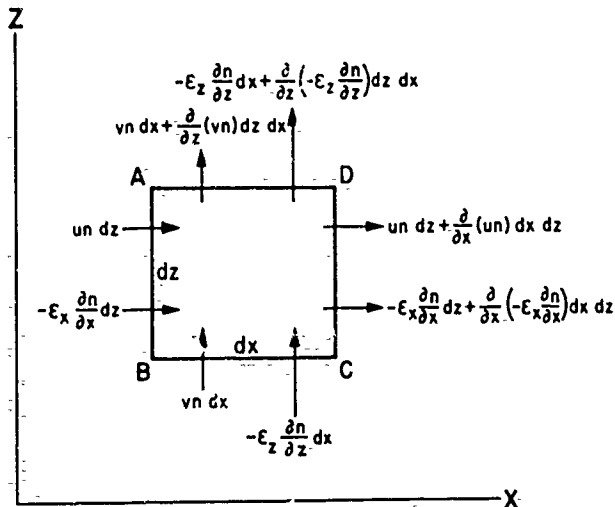


Figure 4. Turbulent diffusion. Element of the longitudinal vertical plane. (Concentration gradients $\partial n/\partial x$ and $\partial n/\partial z$ are negative, i. e., n decreases as x and z increase.)

1. The mass transfer coefficient ϵ_z can be replaced by the eddy viscosity for turbulent air K .
2. The vertical particle velocity v is constant for all levels and equal to $-\underline{w}$, where \underline{w} is still-air terminal fall velocity (this implies that there is no change of particle size with height).
3. The horizontal particle velocity u is equal to the wind velocity, i. e., "slip" is neglected.

Accepting these assumptions and taking the eddy diffusivity in the form given by eq 7, the solution of eq 10 can be written

$$\frac{n}{n_1} = \left(\frac{z}{z_1} \right)^{-w/ku_*} \quad (11)$$

where the subscript "1" denotes a reference level.

Size and fall velocity of blown snow particles

When blown by strong winds, snow crystals are broken and abraded into roughly equidimensional grains with rounded or subangular corners. Particles occur in greatest numbers in the size* range 20μ to 400μ .

Few detailed studies have been made on size and fall velocity of snow particles, although there are a number of brief references in the literature (Table II).

Some of the entries in Table II are for snow types which rarely occur as blowing snow; as already mentioned, snow blown by strong winds usually consists of equant particles with a size of the order of 0.1 mm. Before discussing fall velocity further, more information on particle size is required.

*Unless otherwise specified "size" is used to mean the effective diameter, D_e , defined as $\sqrt{\text{length} \times \text{breadth}}$, in the plane of measurement. Another dimension used in the literature is the projective diameter, or diameter of the circle of equal area; this is defined as $2\sqrt{\text{length} \times \text{breadth}/\pi}$.

Table II. Terminal fall velocities for grains and crystals of snow.

<u>Snow type</u>	<u>Size</u> mm	<u>Mass</u> mg	<u>Fall velocity</u> cm/sec	<u>Reference</u>
Needle	1.5	0.004	50	Mason ⁹³
Plane dendrite	3.0	0.04	30	Mason ⁹³
Spatial dendrite	4.0	0.15	60	Mason ⁹³
Powder snow	2.0	0.06	50	Mason ⁹³
Rimed crystals	2.5	0.18	100	Mason ⁹³
Graupel	2.0	0.80	180	Mason ⁹³
Rounded grains	1.0		200	Gerdel and Strom ³³
Blown particles	0.1-0.5		30-100	Mellor and Radok ⁶²
Blown particles	0.1-0.4		30	Loewe ⁵⁵
Powder snow			50	Nakaya ⁶⁴
Blown snow	0.07		40	Lister ⁵⁴

Careful measurements of size grading in blown snow were made by Diuin²³, Lister⁵⁴, and Budd, Dingle and Radok¹². Diuin's method of sieve analysis seems to have some drawbacks, but it probably gives a reasonably good representation of the actual gradings (Fig. 5). Lister's results (Fig. 6), obtained by measuring and counting grains under a microscope, give particle size in terms of cross-sectional area (i. e., length x breadth of particle). In terms of effective diameter, D_e , they give about 70μ as the most frequent particle size. Budd *et al.* (Fig. 7) show a slow decrease of mean particle size with height above the 10-cm level, at which level the mean particle size is about 100μ . Particles collected 3 cm above the surface were significantly larger than those measured at higher levels.

Budd¹¹ tested histograms of logarithmic particle size for a given level against normal, log-normal, and gamma frequency distributions. Both log-normal and gamma distributions gave adequate representations, but the gamma variate was chosen for analytical convenience.

The mass distribution with particle size is not, in general, given directly by the frequency distribution. Since particle mass is proportional to the third power of particle radius, larger particles make a disproportionately great contribution to the total mass.

For gamma distribution Budd¹¹ gives the mean mass of the particles at any level as

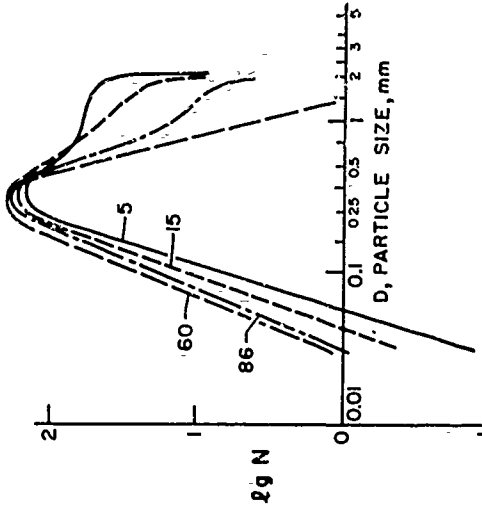
$$\bar{M} = \bar{m} (1 + 3/a + 2/a^2) \quad (12)$$

where \bar{M} is mean mass, \bar{m} is the mass of the particle of mean diameter, and a is one parameter of the gamma variate, * approximately equal to 15 for the Byrd data (so that $\bar{M} = 1.21 \bar{m}$).

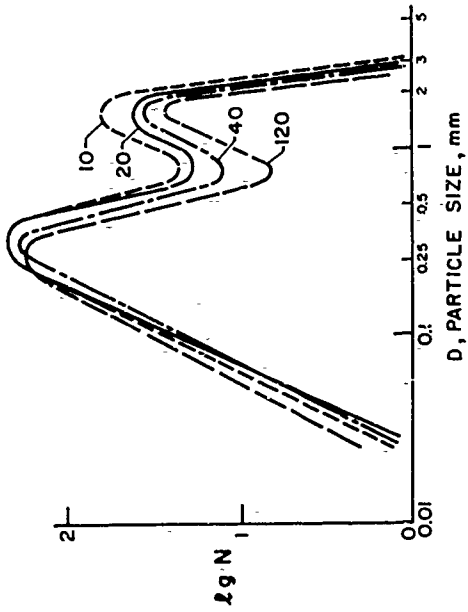
With the preceding information on particle size, the question of fall velocity can be reopened, making particular reference to the relation between particle size and fall velocity. Terminal settling velocities for dispersed aerosols are related to particle size in various forms according to the Reynolds' number. Very small

*In the notation used by Budd, a is the square of the coefficient of variation of the gamma distribution; $a = (\text{standard deviation/mean})^2$.

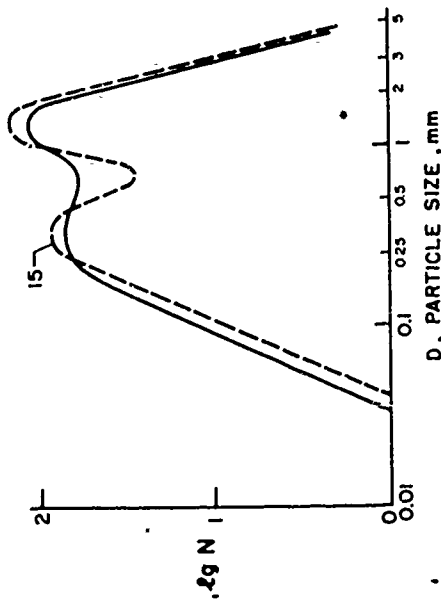
BLOWING SNOW



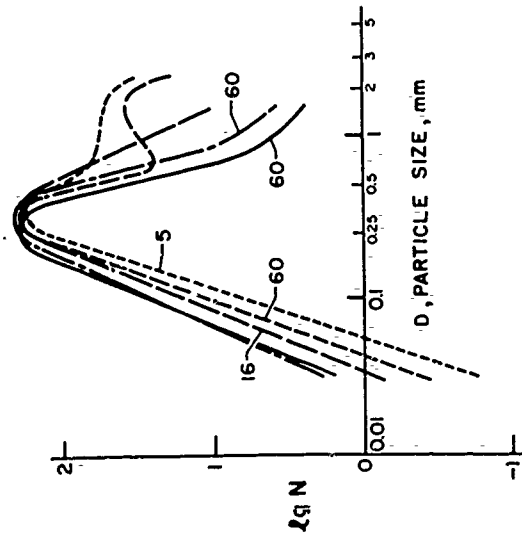
(c) Fresh plate snow. Illustrated is the increasing uniformity due to the mechanical destruction of the particles during screening. Temperatures from -16C to -18C.



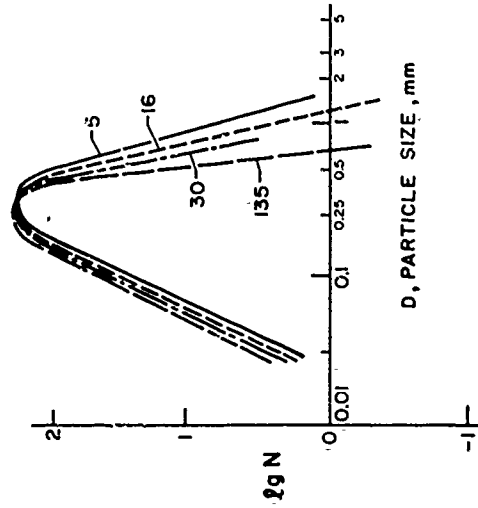
(b) Fresh, stellate snow at low temperatures (from -22C to -26C).



(a) Solid line represents the distorted granulometric composition of fresh, stellate snow; dashed line represents the corrected granulometric composition of the same snow. Temp. -12C.



(d) Snow at the start of a blizzard; temperatures -12C to -21C; size, 0.29 mm.



(e) Snow in a long blizzard; temperature 13 to -19C; average particle size, 0.3 mm.

Figure 5. Size grading by sieve analysis for various snow types, after Diimin²³. Numbers against the curves indicate duration of sieving in minutes.

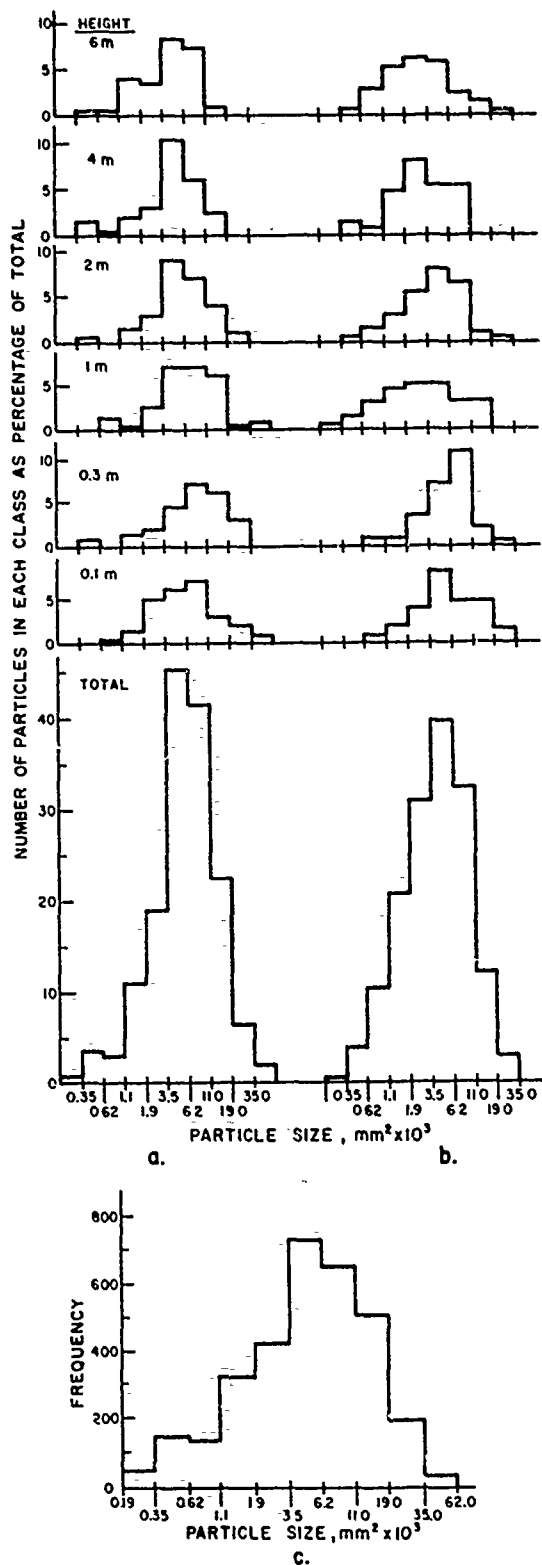


Figure 6. Histograms for drift snow particle size distribution at various heights obtained at Southice, Antarctica by Lister⁵⁴. a and b are two separate samplings in which 50 grains were counted at each level. c is based on a count of 4000 particles sampled in the height interval 10 to 500 cm.

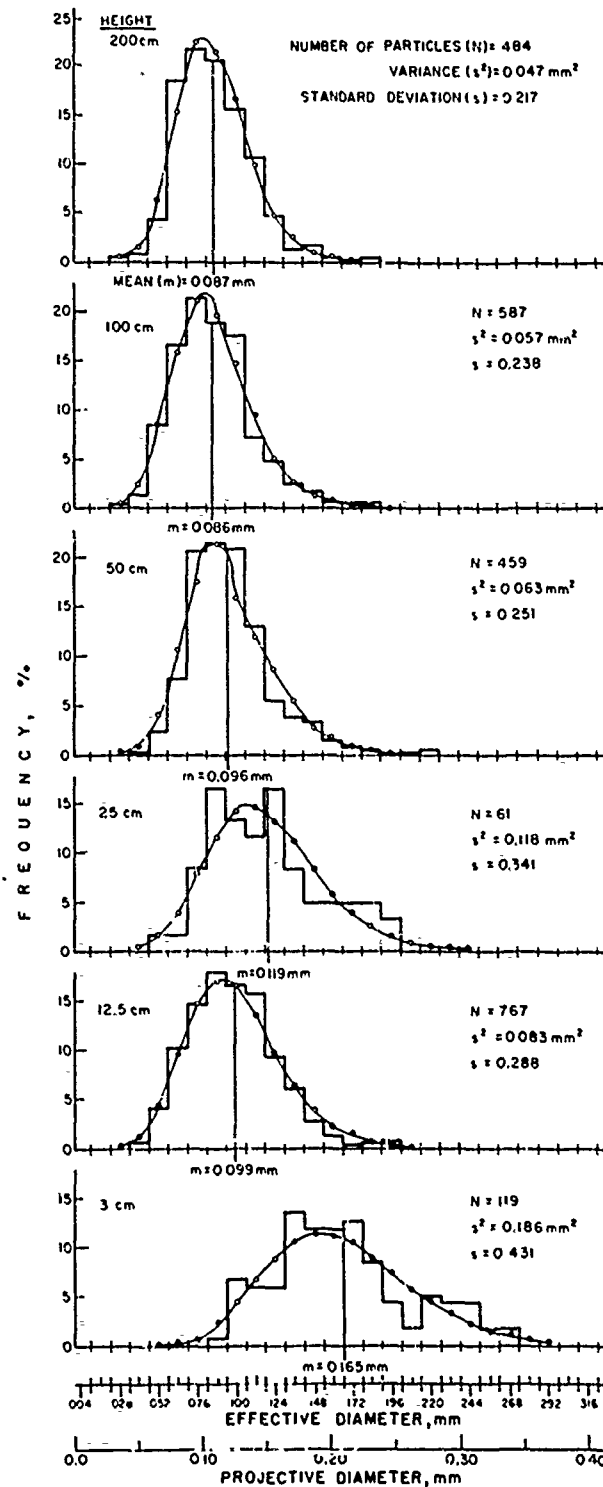


Figure 7. Histograms and frequency distribution curves for particle size in blowing snow. Measurements were made at various heights above the surface at Byrd Station, Antarctica. (From Budd, Dingle and Radok¹²).

particles (laminar flow) obey Stokes' Law, i. e., terminal velocity V_t is proportional to the square of diameter (d) for given densities of air and particle:

$$V_t = C_1 \cdot d^2 \quad (13)$$

where C_1 is a shape factor. The upper limit of size for applicability of Stokes' Law for ice particles in cold air is about 60 to 70 μ . The lower limit, below which Brownian motion becomes significant, is probably less than 1 μ , and therefore of academic interest only.*

Gravity settlement of snow particles ranging in size from about 100 μ to 1 mm corresponds to flow which is transitional between laminar and fully turbulent. In this range aerosols other than snow fall at speeds proportional to the diameter (see, for example, Van der Hoeven⁴¹). Fall velocities for snow particles in still air, measured by the author in a large coldroom at -27C, can also be adequately represented by linear relations:

Angular grains ($D_e = 100\mu$ to 1.5 mm)

$$V_t = 166 D_e \quad (14)$$

where fall velocity V_t is in cm/sec and effective diameter D_e is in mm.

Subangular grains ($D_e = 100\mu$ to 1.5 mm)

$$V_t = 191 D_e \quad (15)$$

Rounded and subangular grains ($D_e = 100\mu$ to 1.5 mm)

$$V_t = 223 D_e \quad (16)$$

With particles bigger than 1.5 mm, flow becomes turbulent and settlement velocity is probably proportional to the square root of particle size.† For a detailed discussion of the settling of single particles and clouds of particles, see Fuchs³⁰.

Gauging blowing snow

Measurements of horizontal transport during snowstorms have been made from time to time for over 80 years, but the early measurements were made only at a single level. Until recently metering was based on mechanical methods involving extraction of snow grains from the air stream, but now attempts are being made to develop more convenient electromagnetic techniques based on scattering and absorption.

Mechanical metering. The usual procedure for measuring mass flow rate has been to expose an orifice to the flow for a timed interval, trapping the suspended snow particles in a container behind the orifice and finally weighing the catch. The devices employed have ranged from simple cylinders open at the upstream end to streamlined collectors with internal arrangements for extracting snow from the air passing through them (Budd, Dingle, and Radok⁴², Garcia³¹, Lister⁵⁴, Mellor⁶⁰).

The objective in mechanical metering ought to be to sample the flow without disturbing it, pass a representative filament through the gauge without impedance, and extract the suspended snow from that filament completely before exhausting the snow-free air. Most gauges have failed to achieve these aims because of one or more of

*It does represent a limit for geometric scaling of particles in wind tunnel tests.

† The foregoing suggests that a general relation between fall velocity V and particle diameter D might have the form $V = V_\infty \exp(-D_0/D)$, where V_∞ is fall velocity for large diameters and D_0 is a constant.

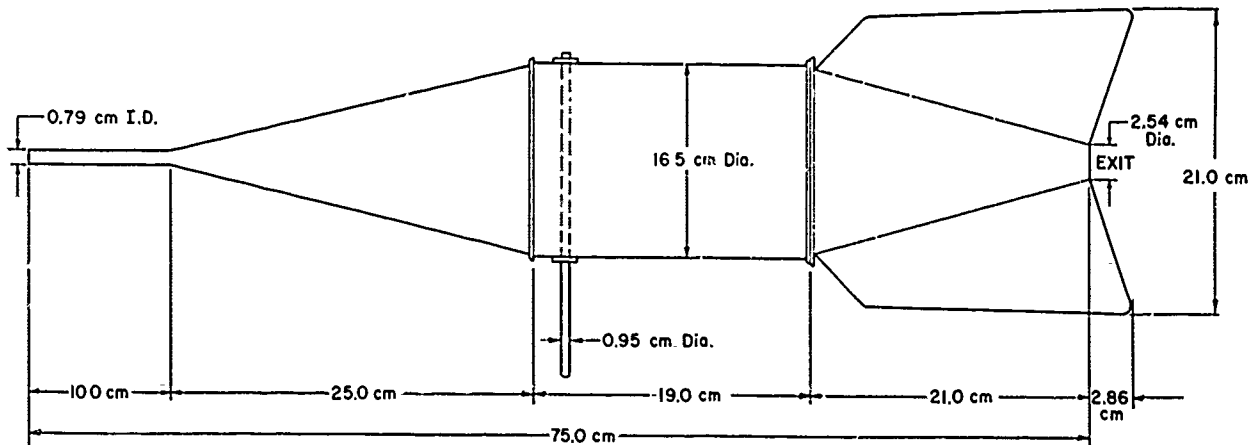


Figure 8. Rocket-type gauge (Mk III) for mechanical metering of horizontal mass flux. A set of these gauges is placed on an anemometer mast to give a vertical profile of mass flux and, from wind velocities measured simultaneously at the gauge levels, a vertical profile of snowdrift density.

the following shortcomings:

- (a) Bluff form and/or inability to orient in the airstream
- (b) Internal bends, baffles or filters
- (c) Imperfect extracting ability.

The closest approach to the ideal, at least in principle, appears to be the Russian "cyclone" gauge, an industrial dust collector adapted for snow studies (Govorukha and Kirpichev³⁶). * It has a "clean" nozzle extending into the flow, and a spiral internal flow path which gives efficient extraction of solid particles. Internal resistance to flow is overcome by an arrangement for sucking air through the gauge, the correct setting of which is indicated by a manometer. In practice, a replica of this gauge has been found somewhat troublesome and unsuited to vertical profiling, although it holds promise as a calibration control.

At present the most convenient gauge for measuring vertical profiles of mass flux seems to be a self-orienting collector with simple streamlining which extracts snow from the air in an expanded internal flow section, where particles settle as velocity is reduced⁶⁰. Revised dimensions of this gauge are given in Figure 8. It has no internal baffles or filters and flow calibrations have been obtained by wind tunnel tests (Pound⁷¹; Budd et al.¹²). Series tests indicate efficient particle extraction, and the internal snow deposit is stable owing to rapid sintering of the particles. Collection efficiency has been discussed and determined by Budd, Dingle and Radok¹². Several gauges are exposed simultaneously at various heights above the surface, together with anemometers which measure the wind profile. Typical exposure periods are in the range 10 to 60 min. The mass of snow caught is found by repeated weighing of the gauge, or the snow may be extracted as melt water for measurement. The gauge measures mass flow rate \dot{m} , so that concentration \bar{n} is obtained when flow rate is divided by wind speed \bar{u} for the gauge level.

Profile measurements very close to the surface have proved troublesome. Special gauges designed for use in the saltation layer⁶⁰ were not successful, but the trap described above has performed satisfactorily in this layer with minor modification¹².

*The theory of cyclone gauges is reviewed by Fuchs³⁰.

Electromagnetic metering.

Blowing snow restricts visibility and light transmission* so much that at the height of a major blizzard the light of a 60-watt lamp may be cut off from view at a distance of 6 m during the hours of darkness. In recent years there have been a number of attempts to devise metering systems based on the attenuation of radiation passing from a source to a detector through snow-filled air.

Chernigov and Vladimirov¹³ tried gamma ray equipment, but found it unsatisfactory. Godshall³⁴ suggested that beta rays are preferable to alpha or gamma rays, but Landon-Smith and Woodberry⁵⁰, who considered beta ray attenuation, discarded the idea on the grounds that air density changes would produce effects comparable in magnitude to those produced by snow particle concentration. Applications of nuclear radiations for drift metering are currently being studied at USA CRREL. Lister⁵⁴ shone a hand lamp on fixed photo cells while snow was blowing during darkness, but practical difficulties prevented him from obtaining useful results. (The writer was similarly frustrated in using a fixed light and a hand-held exposure meter during Antarctic blizzards.) Orlov⁷⁰ described the development of a gauge which responds to attenuation of visible light as snow blows between a collimated lamp and a photo cell; the method apparently proved successful. Landon-Smith and Woodberry⁵⁰ review some unsatisfactory arrangements (including a modulated light beamed at photo transistors) which led up to an operationally successful gauge similar in principle to that described by Orlov.

The final design recommended by Landon-Smith and Woodberry utilizes a projector bulb with constant current supply as the light source, and a cadmium sulphide photoconductor as the measuring detector. A second photoconductor, which is illuminated through snow-free air, provides a reference. The two photoconductors are connected in a balanced bridge circuit. When blowing snow reduces the intensity of light reaching the measuring detector the out-of-balance current is amplified and fed to a recorder. The whole arrangement operates from 12-volt batteries. This meter is calibrated against a mechanical gauge of the type shown in Figure 8.

A reliable radiation device should be far superior to mechanical meters, since continuous records are produced and there is no need for men to go out into the blizzard. There are, however, some calibration problems.

For routine reporting of blizzard density, it seems likely that estimation of visual range could provide the basis of a "daylight" method. The relevant principles are discussed later.

Concentration and mass flux of blown snow

Owing to gauge inadequacies, there is some question about the absolute accuracy of blowing snow measurements quoted in the literature, although relative magnitudes in a vertical profile may be given adequately. The most reliable data seem to be those derived from Australian studies in Antarctica†. ^{12, 19, 62}

Concentration and mass flux as functions of height. Figure 9 shows typical distributions of horizontal mass flux q with height z for strong wind conditions. Above the 1-m level q decreases with increasing height very slowly although it must eventually become zero. Below 1 m, q increases sharply as the surface is approached, reaching a relatively high finite value where z is virtually zero.

The foregoing simple theory suggests that both snow concentration n and mass flux q may be related to height z by an inverse power function. Equation 11 is the required expression relating n and z and, taking the inconsistent but admissible

*See p. 66 for relevant theory.

†Radok, personal communication.

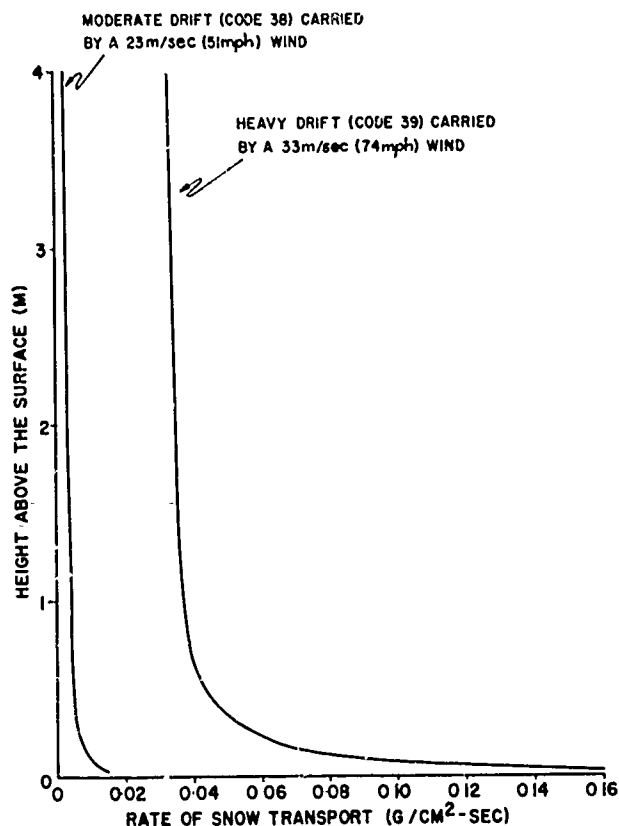


Figure 9. Typical profiles of horizontal mass flux displayed on linear scales.

In the international numerical code for weather reporting, blowing snow conditions are denoted by one of the following:
 36-light to moderate, below eye level
 37-heavy, below eye level
 38-light to moderate, above eye level
 39-heavy, above eye level

assumption of a power law wind profile, a similar relation should apply to mass flux q . In the power wind profile, velocity is proportional to height raised to the power $1/p$, where $p \approx 7$ for neutral stability; thus if $w/ku_* > 1/7$ the flux relationship is of the same form as eq 11:

$$\frac{q}{q_1} = \frac{nu}{n_1 u_1} = \left(\frac{z}{z_1} \right)^{\left(\frac{-w}{ku_*} + \frac{1}{p} \right)} \quad (17)$$

Data from Mawson and Wilkes stations, Antarctica, were plotted on logarithmic scales to test the validity of eq 11 and 17, the variables being normalized with respect to measurements at the 1-m level. It was found that in most cases logarithmic plots were not linear, but curved in such a way that the numerical value of the exponents for eq 11 and 17 decreased with increasing height. Figures A1 and A2 in Appendix A show limits of the domains occupied by \underline{n} and \underline{q} data for Mawson and Wilkes.

BLOWING SNOW

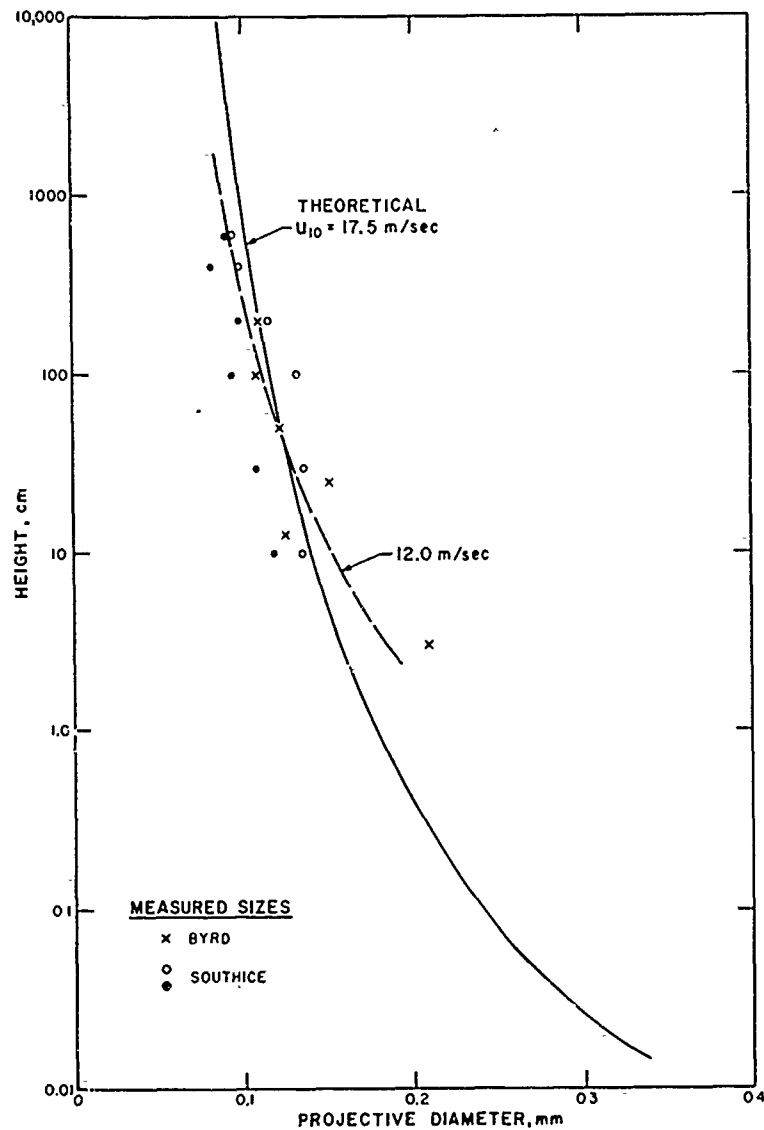


Figure 10. Variation of mean snow particle size with height above surface. (From Budd¹¹).

For any one set of data the wind profile must be considered constant under the steady-state assumption, so that one implication of the curvature found in the logarithmic plots might be a decrease of particle fall velocity with height, i. e., a decrease of particle size with increasing height. Budd¹¹ shows theoretically, with observational confirmation, that changes of particle size and fall velocity with height explain most of the discrepancy between measured distributions of blowing snow and those predicted by the simple theory. Figures 10 and 11 summarize Budd's findings. It also seems possible that saltation in the lowest layers might make concentration and mass flux greater than would be predicted from theory based solely on turbulent diffusion without complete consideration of boundary conditions.

As wind speeds increase, logarithmic plots of n and q against z become flatter, i. e., the snow particles are more uniformly diffused with height. In terms of the simplified theory this means that the magnitude of w/u_* decreases with increasing wind speed, an effect discussed further below.

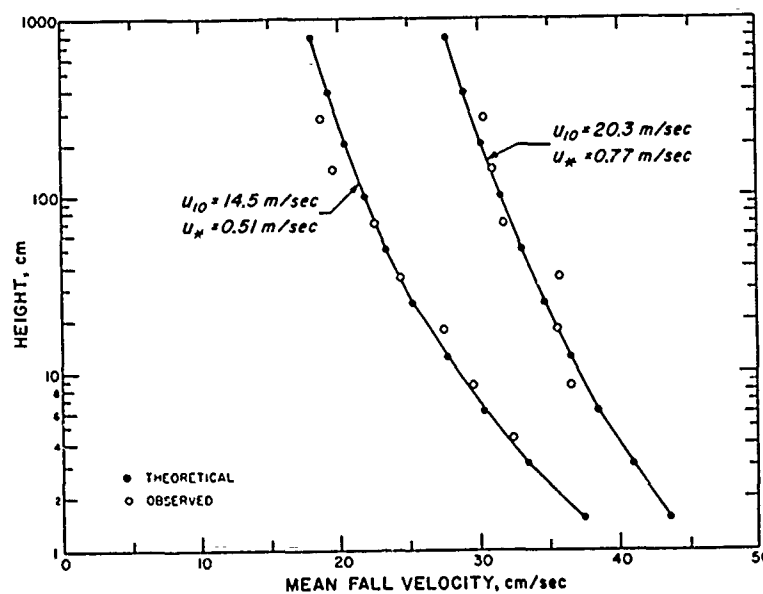


Figure 11. Variation of mean particle fall velocity with height above surface. (From Budd¹¹).

Using a large volume of more recent data, Budd et al.¹² have shown quite definitely that the $\log n$ versus $\log z$ relation is non-linear, at least at the lower wind speeds. Figure 12 gives a selection of their data for a range of wind speeds; the change of general slope with wind speed shows that the snow particles become more uniformly diffused in height as velocity increases. The curvature of the plots, which is greatest at the lowest wind speeds, is apparently due largely to decrease of particle size, and hence fall velocity w , with increasing height. Figure 13 gives a corresponding plot for the mass flux q as a function of height.

Taking an empirical approach, the Mawson and Wilkes data were tested against certain relationships which fit the boundary conditions described for Figure 9 above. Neither the simple exponential decay function nor exponential combinations (hyperbolic functions) represented the data satisfactorily.

Concentration and mass flux as functions of wind speed. From wind profile and visibility measurements during blowing snow conditions, Liljequist⁵³ deduced that drift density, or n , should be proportional to the 5th power of wind speed, as measured at the standard anemometer height of 10 m, although he had no direct measurements to check this deduction. Lister⁵⁴ related his measurements of mass flux q with wind speed at the observation height u , testing both power law and logarithmic forms. The logarithmic relation gave slightly higher correlation coefficients than the power law, and a linear relation between $\log q$ and u was therefore adopted. As another exercise in empiricism, the author has taken Dingle's data from Wilkes Station* and plotted q and n against the wind speed at anemometer height u_0 on logarithmic scales, although this is clearly inconsistent with eq 11. Since the data are not readily available, the plots are shown in Figures A3 and A4 of Appendix A. For gauge heights above 0.5 m, the data can be reasonably well represented by simple power relations, † viz.

$$n = A u_{10}^6 \quad (18)$$

$$q = B u_{10}^7 \quad (19)$$

*Radok, personal communication.

†As a rule, high exponents such as these give a warning that the power function is inadequate.

BLOWING SNOW

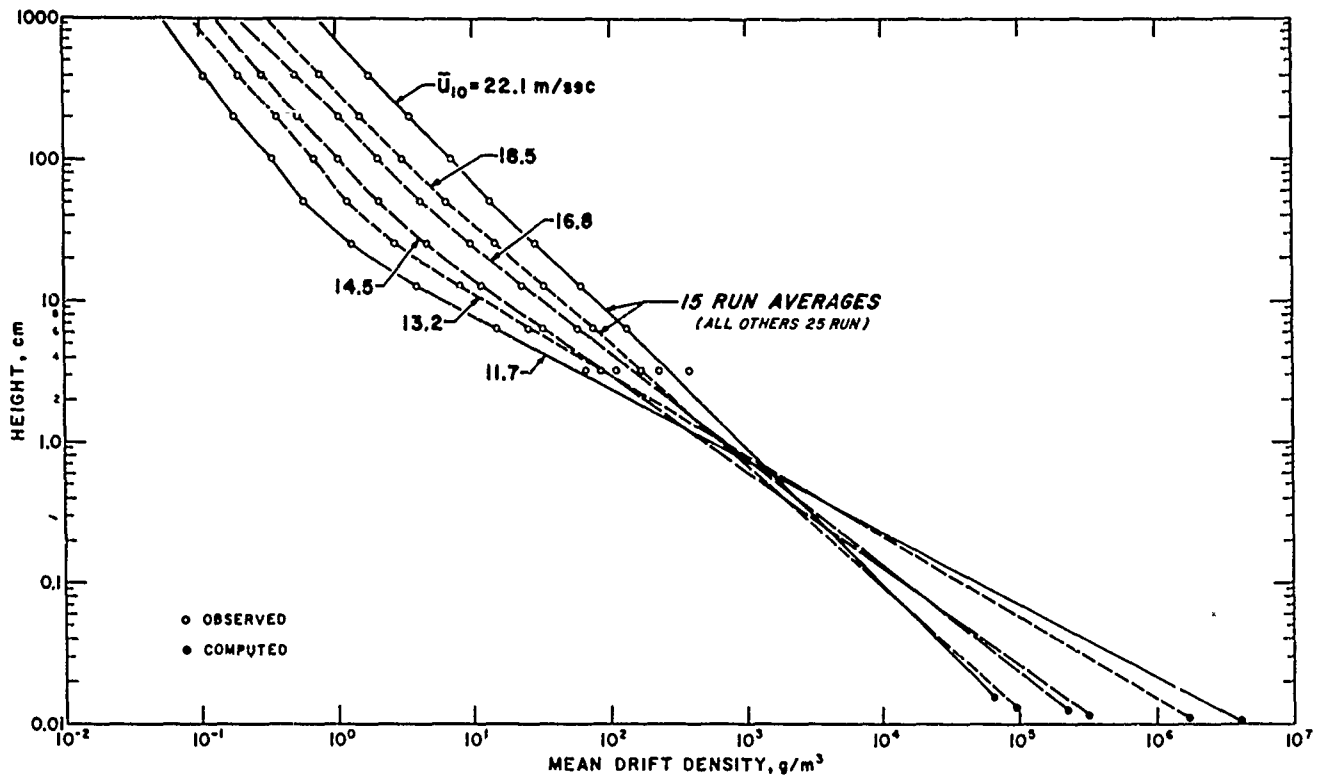


Figure 12. Mean drift density profiles for a range of wind speeds. Lines terminate at mean roughness height z_0 and drift density ρz_0 for the wind speed in question. (From Budd, Dingle and Radok¹²).

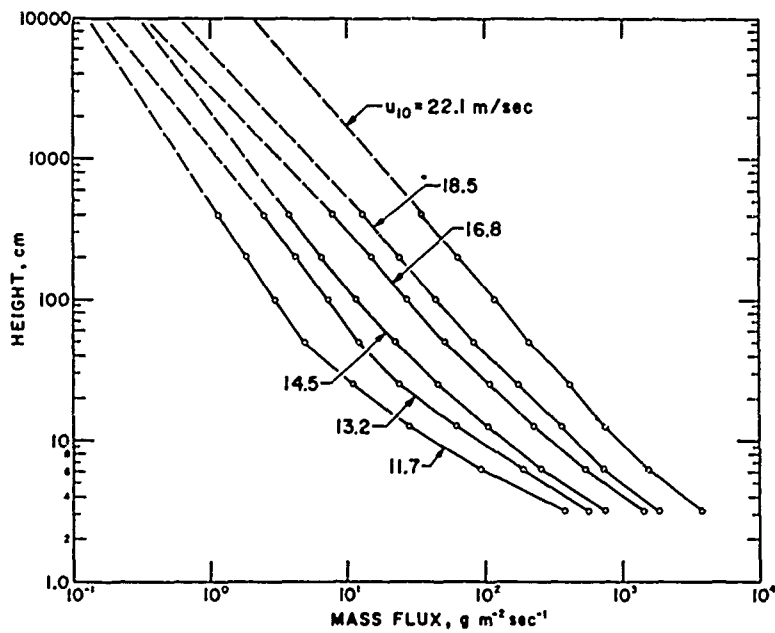


Figure 13. Mean profiles of horizontal mass flux for a range of wind speeds. (From Budd, Dingle and Radok¹²).

where the coefficients \bar{A} and \bar{B} are functions of height. These relationships are displayed on linear scales in Figures 14 and 15. At all gauge levels, \bar{n} and \bar{q} are relatively insignificant for wind speeds less than 10 m/sec; they begin to increase sharply as wind speed rises above 12 m/sec. As mentioned earlier, wind flow over a snowfield becomes fully turbulent when the velocity exceeds about 10 m/sec.

Below the 0.5-m level the data of Figures A3 and A4 suggest some change in behavior. At the higher levels (0.5 to 4.0 m), \bar{q} varies by a factor of 60 to 70 over the range of wind speeds observed, whereas in the lowest layers (3 to 50 cm) the corresponding variation is less—a factor of 20–25. In other words, \bar{q} is less dependent on wind speed near the surface, a fact which is brought out more clearly below. The exponent of a power relation for \bar{q} and u_{10} in the lower layers is 5.4. A similar situation prevails for the variation of \bar{n} with u_{10} ; in this case the exponent is about 4.8. These values bring to mind Liljequist's⁵³ conclusion that increase of surface friction during blowing snow is proportional to the 5th power of wind speed, and his consequent suggestion that drift density in the lowest layers should also be proportional to the 5th power of wind speed.

Data presented in the form of Figures 14 and 15 have a certain practical convenience, but they have little bearing on the mechanics of turbulent diffusion. Avoiding the empirical approach, Dingle and Radok¹⁹ pursued the simplified diffusion theory to obtain a linear relation between the logarithm of concentration \bar{n} and the reciprocal of wind speed at the same height. This relation follows from eq 11 when a substitution for shear velocity u_* is made from eq 1:

$$\ln \left(\frac{\bar{n}}{\bar{n}_1} \right) = - \left(\frac{\bar{w}}{kz} \right) \cdot \ln \left(\frac{z}{z_0} \right) \cdot \ln \left(\frac{z}{z_1} \right) \cdot \frac{1}{\bar{u}} \quad (20)$$

The subscript "1" refers to reference level. Equation 20 may be modified to make the wind variable u_{10} instead of \bar{u} by an appropriate substitution from eq 1.

Equation 20 represents a linear relation between $\ln \bar{n}$ and u^{-1} only if \bar{w} and z_0 are invariant with \bar{u} (or \bar{n}). It has been shown¹¹ that particle size at any given level, and therefore fall velocity \bar{w} , does vary systematically with wind velocity. Furthermore, there is reason to believe that z_0 (which in practice is difficult to measure precisely) increases somewhat with wind speed¹². Thus it is not too surprising that the original test of eq 20 with data for a limited range of wind speeds¹⁹ was rather unsuccessful. However, with a larger volume of data from Byrd Station the broad validity of the relation now seems to have been confirmed; in Figure 16 the results¹² are shown. It is interesting to note that the (negative) slope of the regression line decreases with decreasing height, implying that the dependence of \bar{n} on \bar{u} declines as the surface is approached. This effect shows also in Figure 12, where the density profiles for different wind speeds converge at approximately $z=1$ cm; the simple theory further predicts that at the surface itself ($z=z_0$) \bar{n} decreases as wind speed (u_{10}) increases. Although it is not of much practical interest at this stage, it may be noticed that when the regression lines of Figure 16 are extrapolated to give an intercept on the $\ln \bar{n}$ axis a limiting concentration for extremely high wind speeds is obtained.

It seems probable that the data of Figure 16 are fairly representative; Budd et al. compared the Byrd data with corresponding results from two other stations (Fig. 17), obtaining broad agreement between stations on the coast, on the continental slopes, and on the inland ice of Antarctica.

Total snow transport rate in blizzard. Contrary to some published opinions, the snow blown along in a major blizzard is not necessarily confined to a shallow surface layer; a cloud of blowing snow in the polar regions may be some 300 m deep. It is important for certain glaciological purposes to have an estimate of the total transport rate \bar{Q} , where $\bar{Q} = \int q \, dz = \int \bar{n} \bar{u} \, dz$ with the surface and the top of the drift cloud as the limits of z . Usually, it is difficult to estimate \bar{Q} because of the small range of sampling heights at a drift gauging installation, and also because of ignorance of the detailed wind velocity distribution in the lowest 300 m of the air.

BLOWING SNOW

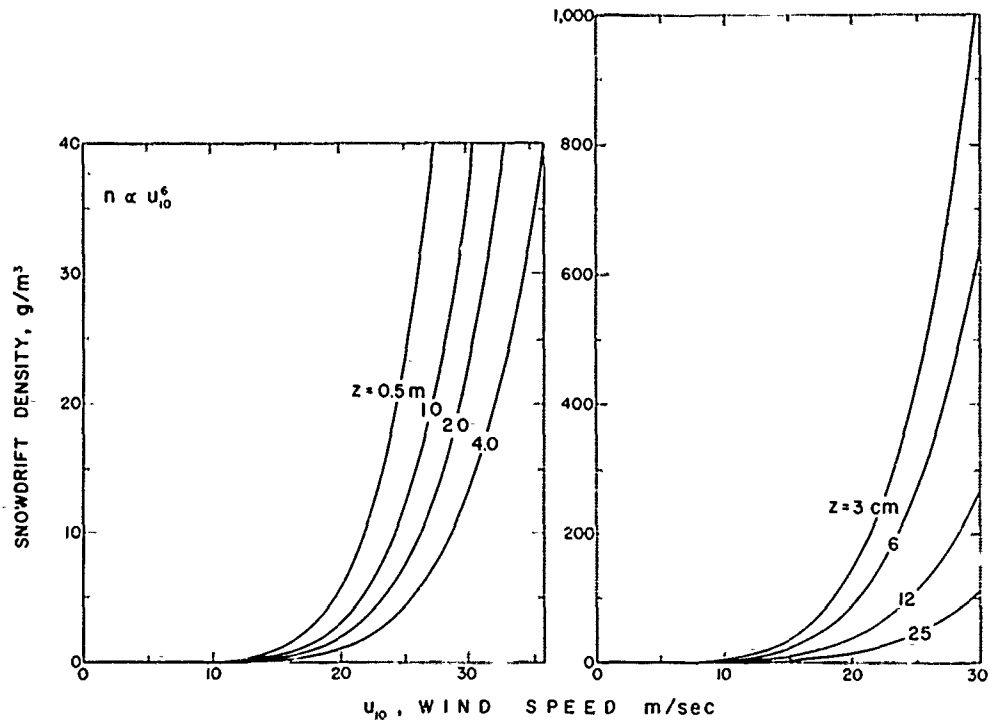


Figure 14. Density of blowing snow as a power function of wind speed for a range of heights. These empirical curves were fitted to data by Dingle and Radok (see Fig. A3) for practical convenience.

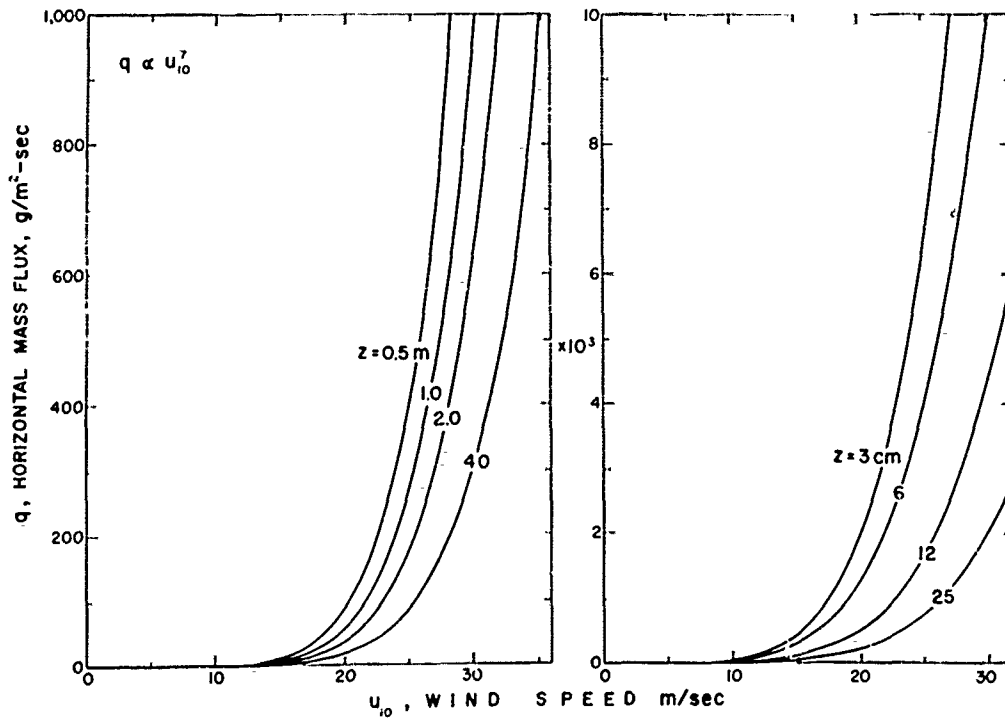
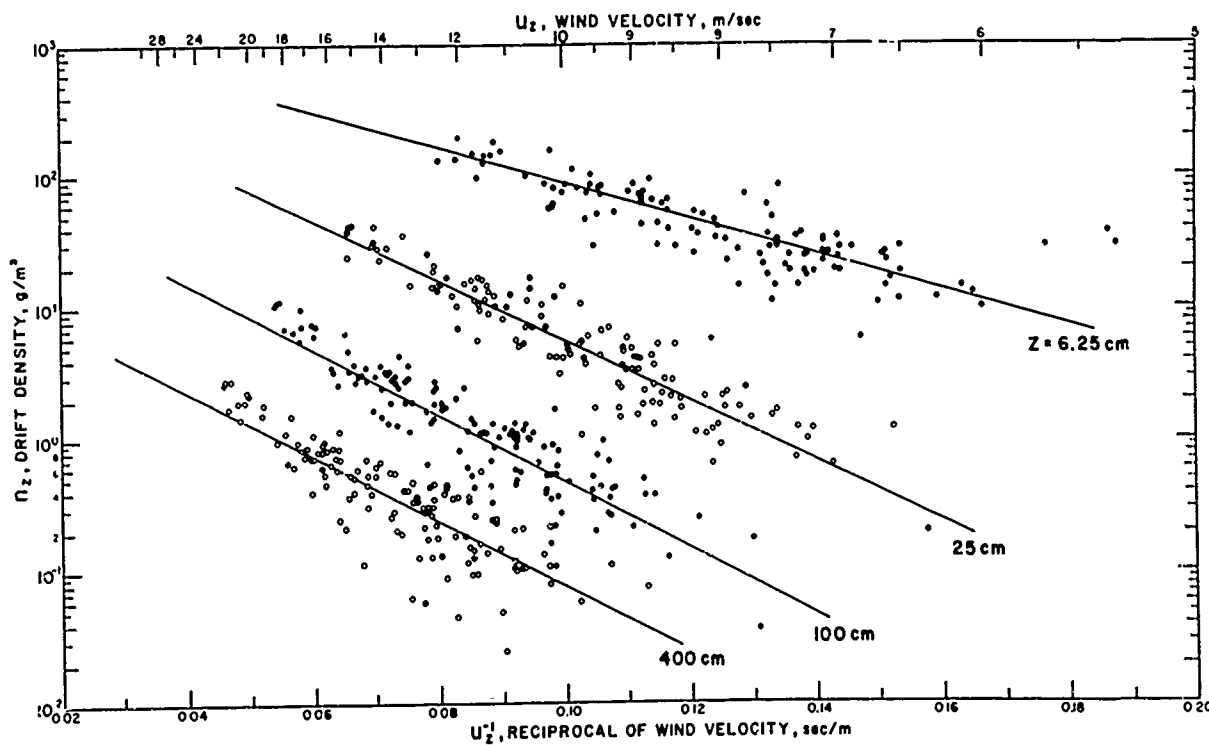
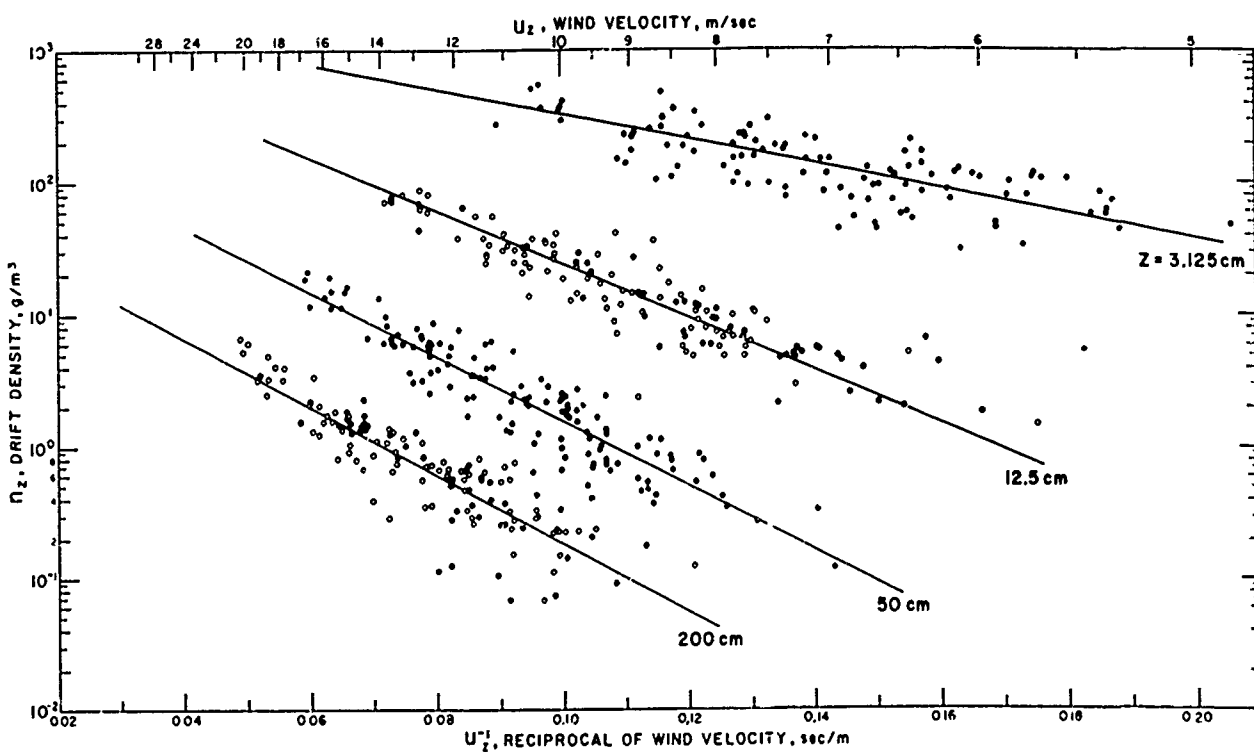


Figure 15. Horizontal mass flux as a power function of wind speed. These empirical curves were fitted to data by Dingle and Radok (see Fig. A4).



(a)



(b)

Figure 16. Drift density as a function of reciprocal wind velocity for a range of heights. (From Budd, Dingle and Radok¹²).

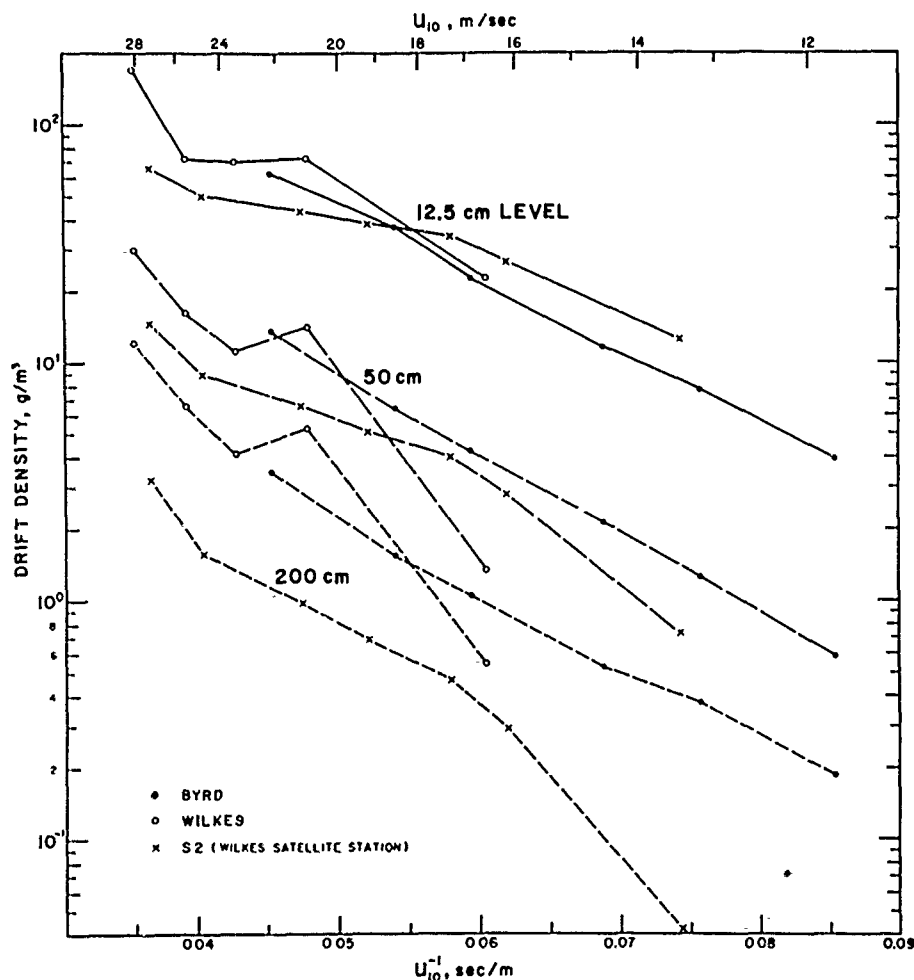


Figure 17. Mean drift density as a function of reciprocal wind velocity compared for three Antarctic stations. Wilkes Station is on the coast, "S2" is on the continental slope, and Byrd is on the inland ice. (From Budd, Dingle and Radok¹²).

Budd et al. measured wind profiles to 300 m by means of "rapid-run" balloon soundings and smoke rockets, and were thus able to extrapolate their surface measurements of \bar{n} and \bar{u} . This led to the relationship between \bar{Q} and u_{10} shown in Figure 18; the graph gives the rate of flow of snow through a normal strip 1 m wide and extending from 1 mm above the surface to 300 m above the surface.

Deposition and erosion

Snow particles are deposited when there is a net flux downward, i. e., when $v_n > \epsilon_z (\partial n / \partial z)$. On an open snowfield deposition commonly occurs because concentration \bar{n} is increased by precipitation from above, or because the vertical mass transfer coefficient ϵ_z decreases as the wind dies down (as barometric pressure gradients decline, or as katabatic* effects dissipate at the foot of a slope). In the vicinity of an obstacle, velocity reduction in the eddy zones may also permit particles to settle. In some cases there is deposition resulting from instantaneous adhesion of particles which contact the surface, usually because of high temperature. (A special case is rapid sintering of very small particles comminuted by snow- or ice-cutting machinery.)

*Katabatic winds are density currents which flow down surface slopes subject to Coriolis deflection—see Ball⁴.

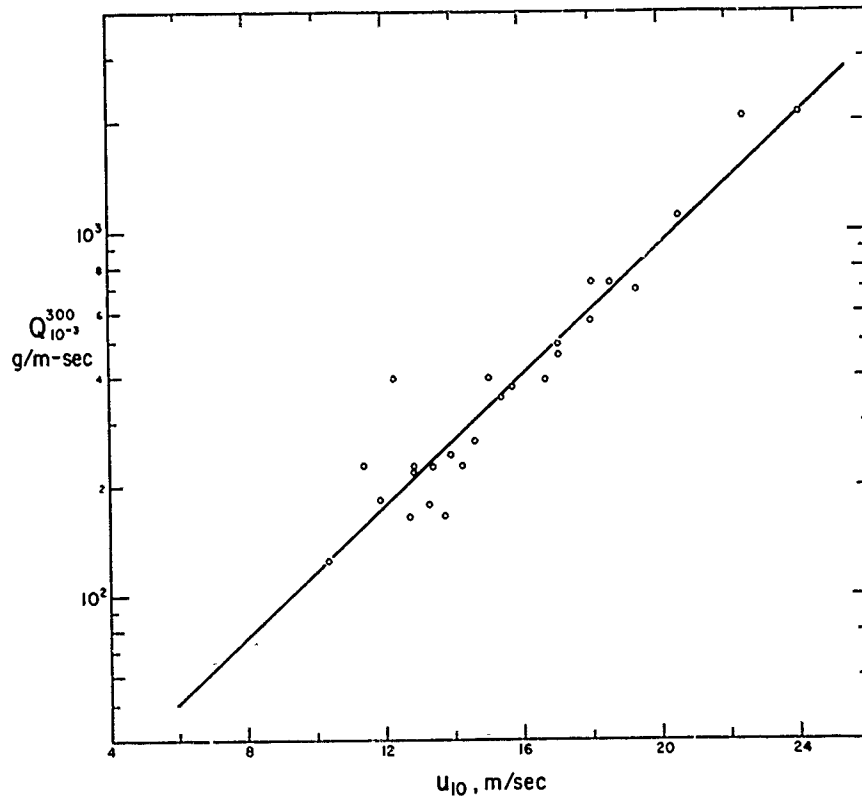


Figure 18. Total snow transport rate as a function of wind speed at Byrd Station. Transport rate \bar{Q} is $\int q dz$, or $\int nu dz$, from $z=1$ mm to $z=300$ m. (From Budd, Dingle and Radok¹²).

As far as is known, "saturation" concentrations of snow particles have not been determined directly, although approximate values can perhaps be estimated from field data. For the Antarctic studies described above, the highest values of n recorded at a given level are probably close to saturation values; if upper limit envelopes are drawn parallel to the fitted lines in Figure A3 it is found that maximum concentrations are about three times greater than the mean values represented by the curves of Figure 14 (left).

During deposition the low frequency pulsations of wind velocity, or gusts, create periodic variations in the intensity of vertical exchange, so that there is a consequent fluctuation in both deposition rate and mean particle size for the deposited material. The intermittent size selection results in a finely banded structure for the deposited snow, with distinct layers of the order of 1 mm thick. This effect perhaps deserves some study, since the fine laminar structure of polar snow may preserve a record of past wind characteristics.

A snow surface is eroded when the wind speed and the surface roughness are sufficient to develop a shear stress great enough to break particles free from the surface. The magnitude of this critical shear stress will vary with the size of the snow grains and with the degree of intergranular bonding in the surface layer. With cold, cohesionless, fine-grained snow, wind speeds of a few meters per second may suffice to dislodge particles, but not to diffuse them into the airstream by turbulent exchange. Under these circumstances the particles will merely roll or bound along the surface in a thin layer, commonly no more than 10 cm thick. Not until turbulence is well developed and $\epsilon_z (\partial n / \partial z) > vn$ will particles be carried up to and above eye level. The foregoing summary of Antarctic data suggests that wind speed has to exceed 10 m/sec for this to happen.

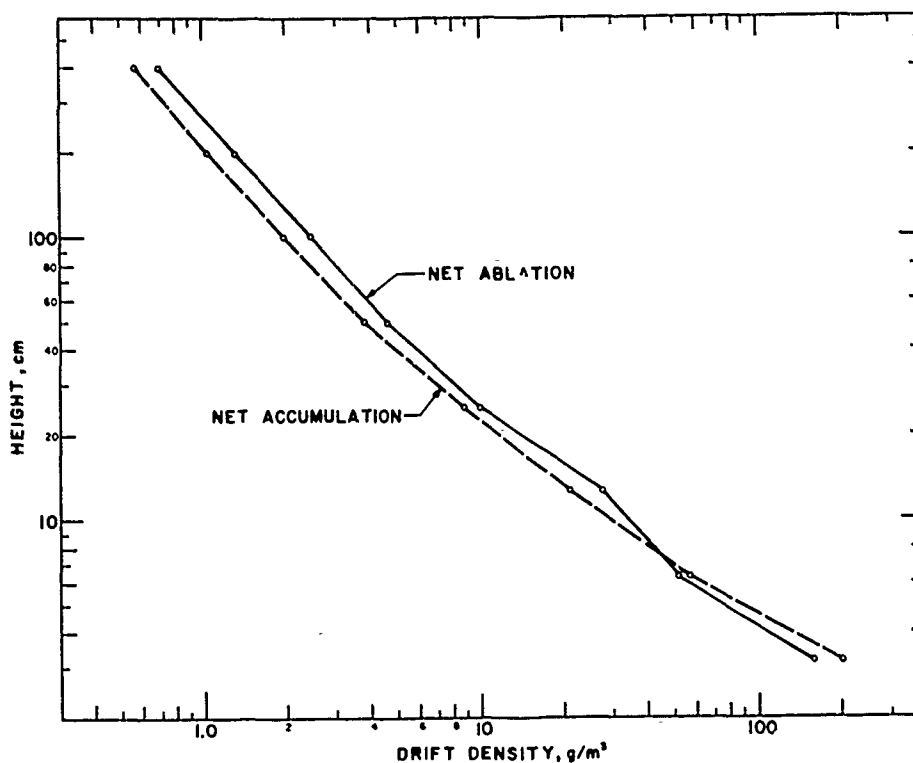


Figure 19. Mean drift density profiles for periods of net surface accumulation and periods of net surface erosion. The averages are made meaningful by matching accumulation and erosion profiles in pairs for corresponding wind conditions. (From Budd, Dingle and Radok¹²).

Budd et al.¹² have attempted to distinguish between drift density profiles for periods of net deposition and periods of net erosion by averaging for matched pairs of data. In Figure 19 they compare mean profiles for net accumulation and net ablation; at levels above 10 cm, concentration is higher during periods of net ablation, while near the surface concentrations are lower during net ablation.

A firmly bonded snow surface, or even an ice surface, can still be eroded slowly by evaporation, which is aided by wind turbulence. Some erosion features produced by evaporation show a marked resemblance to corresponding features produced by particle transport (e. g., wind scoops); this is not too surprising in view of the broad similarity between turbulent transport of small particles and turbulent transport of water vapor.

Preferential deposition and erosion along a snow surface are matters of considerable interest. Variations of deposition rate with distance along a flow line occur for a wide variety of surface perturbations, from gentle undulations with wavelengths of 10 km or more to abrupt obstacles such as snow fences.

Measurements of snow accumulation on polar ice caps have revealed that, in windy areas, a correlation exists between accumulation rate (temporal mean over 1 year or more) and large scale surface topography (undulations with wavelengths in the range 1 - 10 km). Highest accumulation rates are found in the troughs of undulations, and lowest accumulation rates near the crests. Correlation of accumulation rate with surface slope indicates that there is a slight phase shift, however, with maximum accumulation just downwind of the troughs and minimum accumulation just downwind of the crests (Black and Budd⁹⁴). This implies that, neglecting flow of the

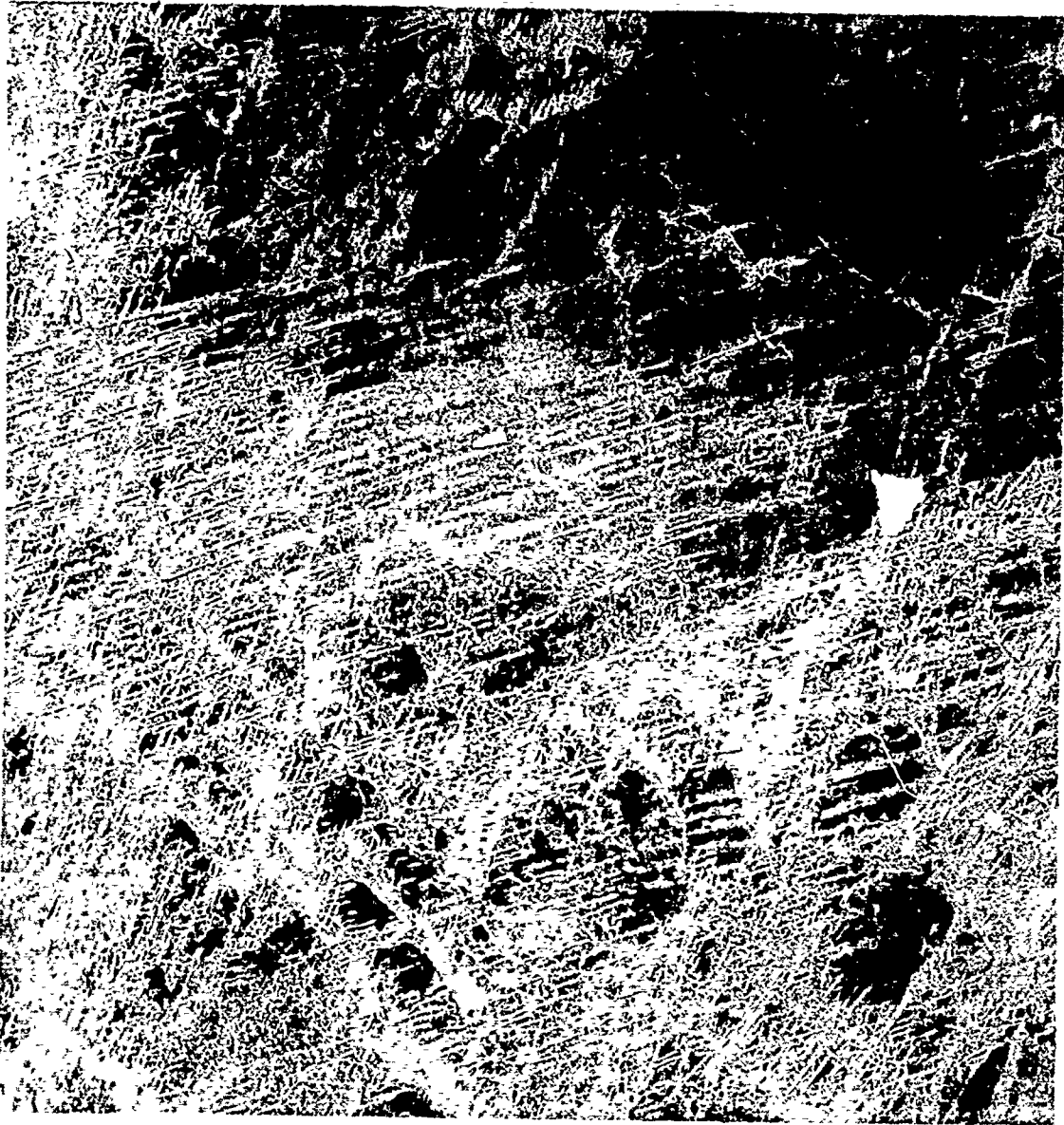


Figure 20. Snow dunes deposited on sea ice. This air photo covers an area 3.3 km square. Winds generally from lower left.

ice sheet, undulations lying athwart the wind direction tend to be displaced upwind. A contrary trend, displacement downwind, has been reported by Dolgushin²⁶, but there is some question as to the nature of the features referred to.

On a much smaller scale, evidence of non-uniform deposition is given by dunes (seifs), sastrugi, barchans, waves, and ripples. These features all have characteristic orientations (dunes and sastrugi aligned with long axes in the wind direction; barchans, waves, and ripples transverse to the wind direction), but on flat snowfields they appear to occur in randomly distributed groups when seen from the air. This probably follows from random distribution of initial surface irregularities. An exception might be made in the case of dunes, which sometimes form along preferred parallel tracks, 10 m or more apart, aligned in the wind direction (Fig. 20). The effect is suggestive of parallel streaks of foam sometimes seen on windswept water.

Dunes are rounded whaleback deposits which form during precipitation-fed blizzards. The long axis of the dune is aligned in the wind direction, and a longitudinal profile takes a streamlined form, the "hump" of the upwind end flaring away downwind (like an airfoil or a drumlin). The lengths of dunes formed on open snowfields range from less than 10 m to 100 m or more; length and plan aspect ratio* seem to increase with the wind speed at time of deposition. Width is roughly one-fifth of the length for typical dunes (≈ 10 m long), and heights vary from about 0.5 to 2 m over the complete size range. Dune migration may occur during the period of deposition (when observation is very difficult), but no bodily movement of dunes subsequent to the originating storm has been reported. At the relatively high temperatures associated with cyclonic storms, deposited snow sinters rapidly, so that bodily migration of the deposit is not likely to occur. Later on, wind action may erode a dune or deposit more snow in its lee, so that it becomes a composite feature. The sharp-edged sculpturing of a dune by katabatic winds may produce giant sastrugi at its edges.

Sastrugi are generally regarded as being the sharp-edged longitudinal ridges which characterize the surface of a windswept snowfield (Fig. 21), although the term was originally used (and sometimes still is) in a less restrictive sense to include any wind-formed irregularity. The shape and distribution of typical "plowshare" sastrugi are suggestive of the boundary layer structure at plane surfaces outlined by Rotta⁷⁹: arrays of elongated "islands of hesitation" break the flow immediately adjacent to the plane surface, and longitudinal screw vortices propagate from the edges of these "islands." The effect has been attributed to interaction between the laminar sub-layer and the turbulent boundary layer. On polar ice caps sastrugi are frequently formed by katabatic winds, i. e., they commonly form when snow is being redistributed by "dry" winds rather than in periods of precipitation. Their long axes are then aligned with the direction of the katabatic wind which formed them, which in general is different from the cyclonic storm wind direction. Typical sastrugi (leaving out of account the giant variety formed in association with dunes) are about 1 to 2 m long, and the greatest width is roughly one-fifth of the length. Common heights (trough to crest) are 10 to 15 cm, but most polar travelers have formed the impression that there is a direct correlation between sastrugi height and wind speed at the time of formation. It might be that plan aspect ratio would provide a better correlation with wind speed. Heights of giant sastrugi are related to the dimensions of the associated dunes, and heights of 1.5 m have been measured. Sastrugi move during the period of formation while the cold snow grains are still unbonded; migration rates in the range 0.1 to 1.0 cm/min have been measured by the author. After age-hardening sastrugi cease to migrate bodily, although they are subject to erosion.

The orientations of both dunes and sastrugi are useful indicators of wind direction in remote regions; sastrugi formed on ice caps by strong katabatic winds give, after analysis, a good idea of surface contours (Mather and Goodspeed⁵⁹, Mather^{57, 58}, Ball⁴) and thereby indicate the direction of ice flow.

Small barchans (crescentic dunes) form under certain conditions. The horns of the crescent point downwind, and the barchan may migrate during the formation period at rates of about 2 cm/min (Benson¹⁰). Typical examples (Fig. 22) measure about 2 m across. Barchans seem to be a transitional form between transverse waves and ripples and longitudinal features (sastrugi).

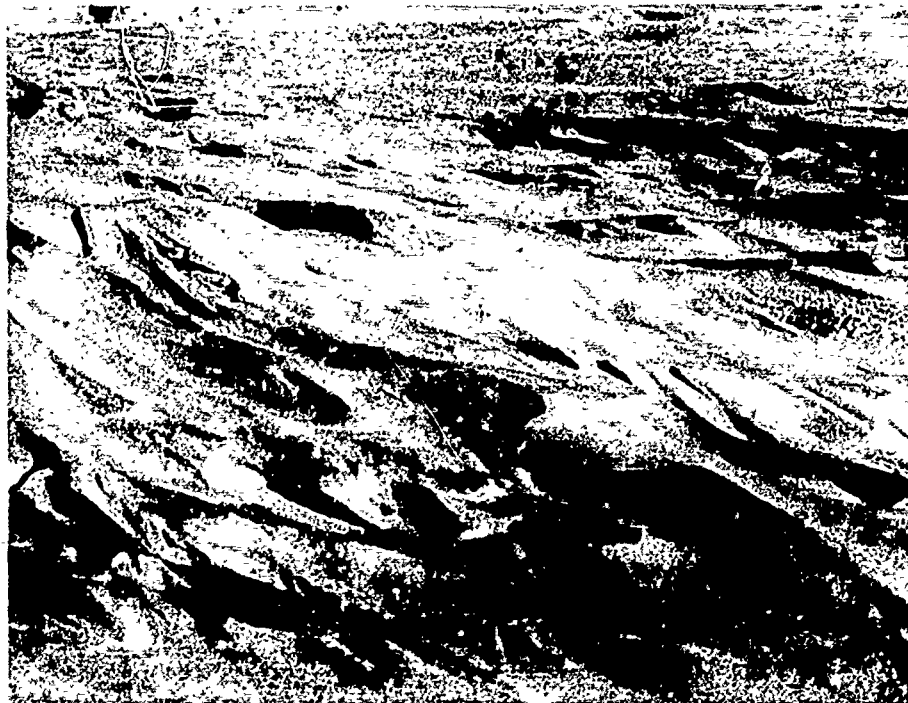
The term waves has been applied to transverse undulations with wavelengths from 5 to 10 m and heights from 10 to 20 cm (Cornish¹⁴). They form with strong winds and low temperatures in certain locations, and travel at about 5 cm/min.

Ripples are small transverse waves which form in cold loose snow where winds are locally light (sometimes in the shelter of obstacles or dunes when brisk winds are blowing). Typical wavelengths are 10 to 40 cm, and wave heights (trough to crest) are about an order of magnitude smaller than the wavelength. Seen in plan, troughs and crests are sinuous (Fig. 23), and as wind speed increases this feature becomes more pronounced, culminating in complete breakdown of the transverse form with transition to a longitudinal structure (sastrugi). Ripples migrate during the formation period at rates of 5 to 20 cm/min.

*Ratio of length to width.



(a)



(b)

Figure 21. Sastrugi. (a) A strongly elongated type.

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(c)



(d)

Figure 21 (Cont'd). Sastrugi. (d) A form closer to transition from transverse ripples to longitudinal sastrugi.



Figure 22. Snow barchan.



Figure 23. Transverse ripples on the snow.



Figure 24. Patches of sharp-edged sastrugi overlying discontinuous "pavements" of pocked hard snow.

Formation of the features mentioned above often involves both erosion and deposition alternating locally. There are, however, other wind-formed irregularities which are pure erosion features. Local intensification of evaporation by turbulent eddying has already been mentioned; there is also pitting and scouring resulting from fracture of surface crusts and mechanical removal of snow grains. There is no accepted terminology for patterns "etched" in crusted and stratified snow surfaces, although words like "pocking" and "pitting" have been applied. This type of erosion frequently occurs on relatively smooth crusted surfaces where successive but discontinuous layers of deposited snow overlap to give an effect which has been described by words such as "pavements," "tiles," and "terraced surfaces" (Fig. 24). "Pocking" or "pitting" appears to occur when the wind shear becomes just sufficient to break a surface crust at its weak points; if the snow under the crust has relatively low cohesion, as is frequently the case, loose grains are plucked out through the breaks in the crust.

The initiation of wind erosion and sastrugi formation can be nicely demonstrated using the propwash of an aircraft. The machine is swung so that the propwash blows over snow which has not been artificially disturbed, and the throttle is opened gradually, so that the surface shear stress slowly increases to the critical value for disturbance of the surface. In deep, loose snow, migrating sastrugi can be formed this way.

The deposition and erosion features described so far are those found on open snowfields, but in the vicinity of large abrupt obstacles flow disturbance is severe and the effects are more pronounced. The deposits laid down near rocks, ridges, trees,

fences, structures, and similar surface obstructions are known as snowdrifts, and their practical importance merits more detailed discussion.

Boundary layer separation and turbulent wakes

The turbulent boundary layer of wind flowing over an open snowfield produces the characteristic but randomly distributed deposition-erosion features mentioned above, probably by repetitive flow separation, as suggested for sastrugi. When the wind encounters an abrupt disturbance of the surface, local separation of the boundary layer occurs on a larger scale. Energy is dissipated by eddy formation near the obstacle, and preferential deposition and scouring occur when snow is borne along by the wind. Snow deposits and accumulates where surface shear is low, but there may actually be erosion in the eddy zone where surface shear due to intense eddies is high.

The phenomenon of separation, which is responsible for the major eddies that form snowdrifts, is familiar to engineers as the cause of stalling on airplane wings and cavitation in ships' propellers and turbine blades; it is also utilized in diffuser tubes. When a fluid flows over a curved surface, or wind blows over the crest of a slope (Fig. 25), in such a way that streamlines diverge, velocity u decreases and pressure p increases with distance x (du/dx is negative, dp/dx is positive). Particles in the boundary layer lose kinetic energy by friction as they travel against the adverse pressure gradient, and eventually the impressed fluid pressure forces a reverse flow at the boundary. The limit of the boundary layer where $u=0$ thus separates from the surface, the point of separation (stagnation point) being that point where the vertical velocity gradient at the surface becomes zero [$(du/dy)_{y=0}=0$]. The eddies created by flow separation form a turbulent wake, in which kinetic energy is dissipated by rotation.

The eddy pattern in the vicinity of an obstacle has a direct bearing on the formation of snowdrifts, but in spite of this there is remarkably little accessible information on the wakes created by fences, steps, and structures projecting into the air-stream. (A surprising lack, for eddies behind a vertical barrier were illustrated almost five centuries ago by Leonardo da Vinci.) The sole study directed specifically towards snowdrifting problems was made by Finney²⁷ in 1934; his modeling technique was unsophisticated by modern standards, but the data have not been superseded.

Figure 26 shows the eddies traced by Finney for model fences at 1/24 linear scale with a 15 mph air stream in a 2 ft x 2 ft wind tunnel. The length of the eddy zone on the lee side of the fence did not change as wind speed was varied from 10 to 30 mph, but there was a direct dependence on fence height (Fig. 27).

Finney also determined the eddy pattern for a 15 mph wind blowing over a model bank of triangular cross section (Fig. 28). He went on to find how the length of the eddy zone varied with the height of the bank and with change in inclination of the right-angled wedge representing the bank (Fig. 29).

Eddy patterns are depicted in the literature for various types of surface disturbance, but usually there is no direct supporting evidence. Figure 30 gives a simplified impression of the eddies caused by steplike disturbances, embankments, and wide buildings, based on a number of presentations in the literature. Separation on the upwind side of obstacles is probably illustrated adequately, but the structure of the lee wakes is certainly oversimplified in some cases. The number of major eddies in a wake, and their locations, are likely to change with wind speed for an obstacle of given size, so that in any fixed location relative to the obstacle the sense of rotation of the eddy may change as wind speed changes. The main eddy on the lee side of a pitched roof building has been reported as changing its direction of rotation during gusting (Goldstein,³⁵ following Nøkkentved⁶⁶). There also seems to be a possibility of backflow separation, leading to counter eddies, when wind blows over a crest and down a slope.

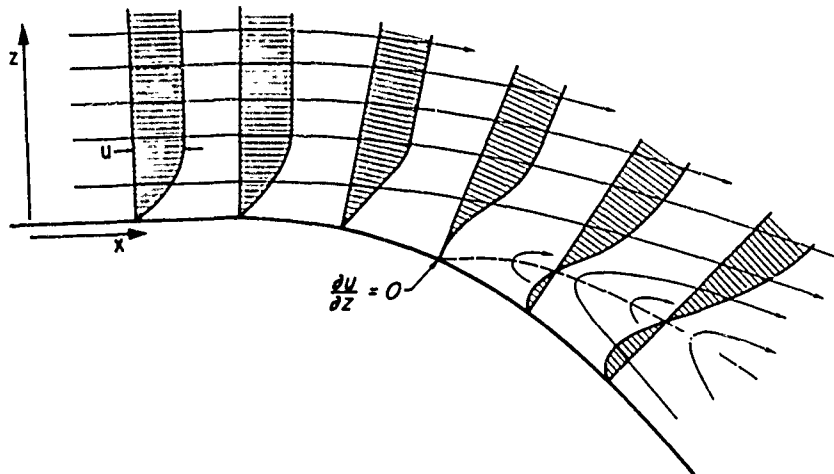


Figure 25. Boundary layer separation.

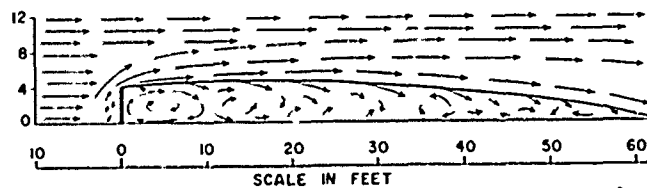
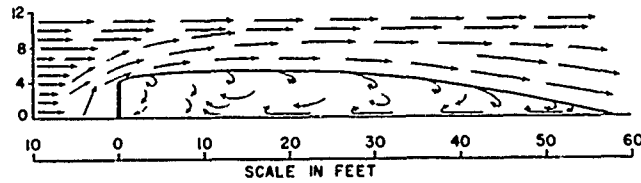
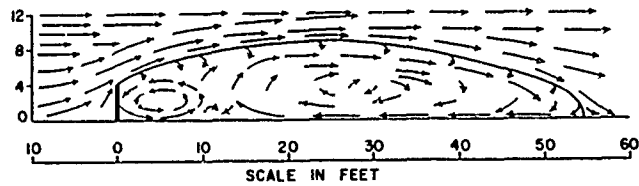


Figure 26. Turbulent eddies formed by long snow fences set normal to the wind direction, from tests by Finney²⁷. (Top) solidly sheeted fence, (middle) horizontal slat fence with 50% density, (bottom) vertical slat fence with 50% density.

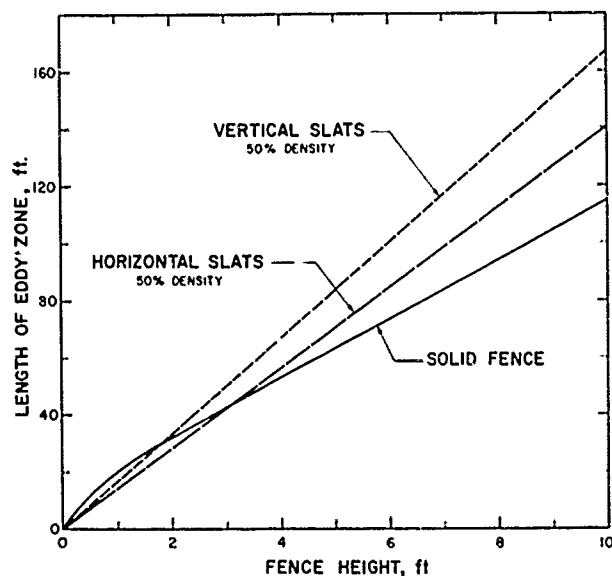


Figure 27. Length of the eddy zone behind snow fences as a function of fence height. (Dimensions scaled up from 1/24-scale models.) (From model tests by Finney²⁷).

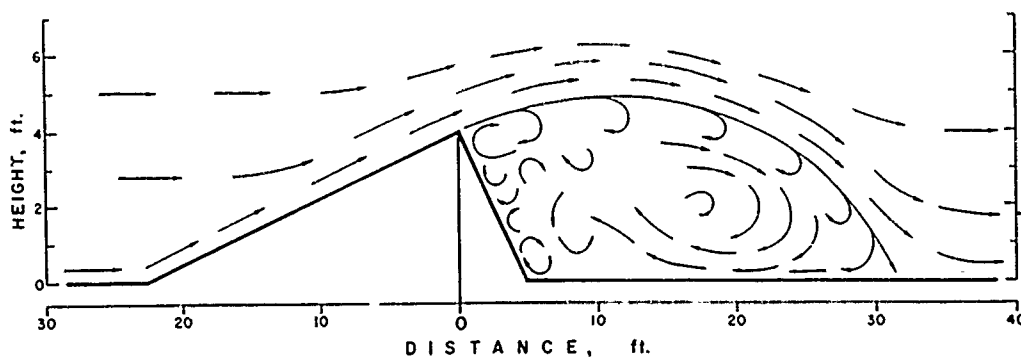


Figure 28. Eddies formed by a bank. (Dimensions scaled up from 1/24-scale model.) (From model tests by Finney²⁷).

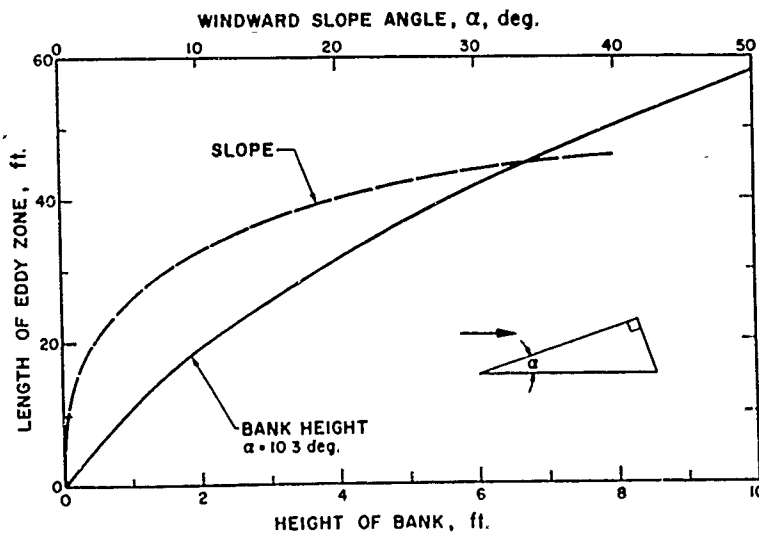


Figure 29. Length of the eddy zone behind a model bank as a function of bank height and windward slope. (Dimensions scaled up from 1/24-scale models.) (From model tests by Finney²⁷).

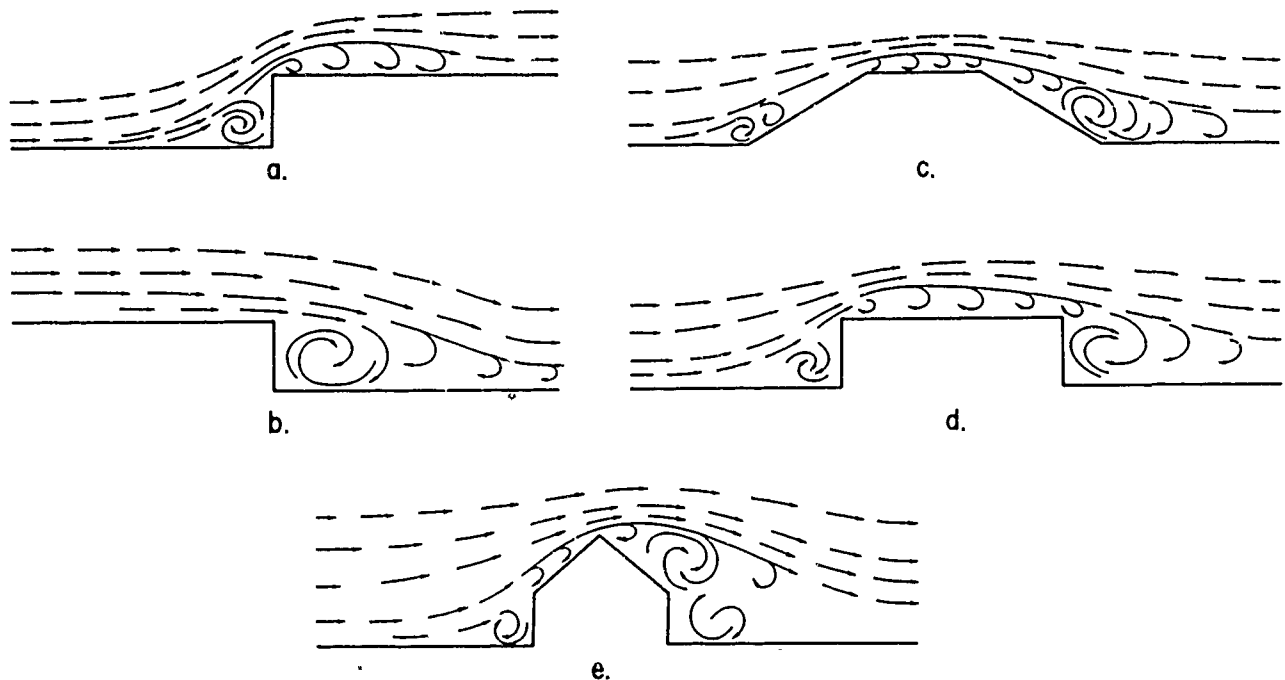


Figure 30. Simplified impression of eddies formed by typical obstructions (not to scale).

Wind tunnel testing

The value of wind tunnel testing in snowdrifting studies has been widely recognized for about 30 years, but the opportunities to engage in model investigations have been limited. In the early experiments there does not appear to have been any detailed model analysis, but useful results* were obtained by simple methods. During the past decade there has been a more rigorous approach and techniques have been refined. There are, however, serious difficulties in designing experiments to conform simultaneously with the many scaling criteria now recognized.

Finney's 1934† tests at Michigan State College were based on geometric scaling of fences and banks, flaked mica and fine sawdust being blown against $1/24$ scale models set on a bed of sandpaper in a 2 ft x 2 ft tunnel. Similar work was done later in Denmark by Nøkkentved⁶⁶ (1940) using fine sawdust. Some wind tunnel measurements were made using powdered gypsum and peat dust by Becker⁸ (1944). In Japan, snowdrifting around buildings was investigated using magnesium carbonate powder as model snow (Kimura and Yoshisaka, 1942)⁴⁵. In none of these tests was much consideration given to kinematic or dynamic scaling.

In 1949 Imai⁴⁰ introduced similitude principles into a theoretical discussion of snowstorm modeling, and recommended magnesium carbonate sieved through a 0.2-mm mesh as a model substitute for snow. Later, Tanifuji and Ogawa⁹⁰ (1956) reported extensive tests on fences and embankments, including wind tunnel tests (using magnesium carbonate) based on some prior analysis. However, it was not until about 1958 that a thorough study of the modeling problem was initiated. Scaling criteria were developed by Kelly, Keitz and Strom⁹⁵ and these were applied in a

*Some of these results are summarized later.

†It is reported that wind tunnel tests were made on model snow fences in 1933 by Moroshkin (Kungurtsev⁴⁹).

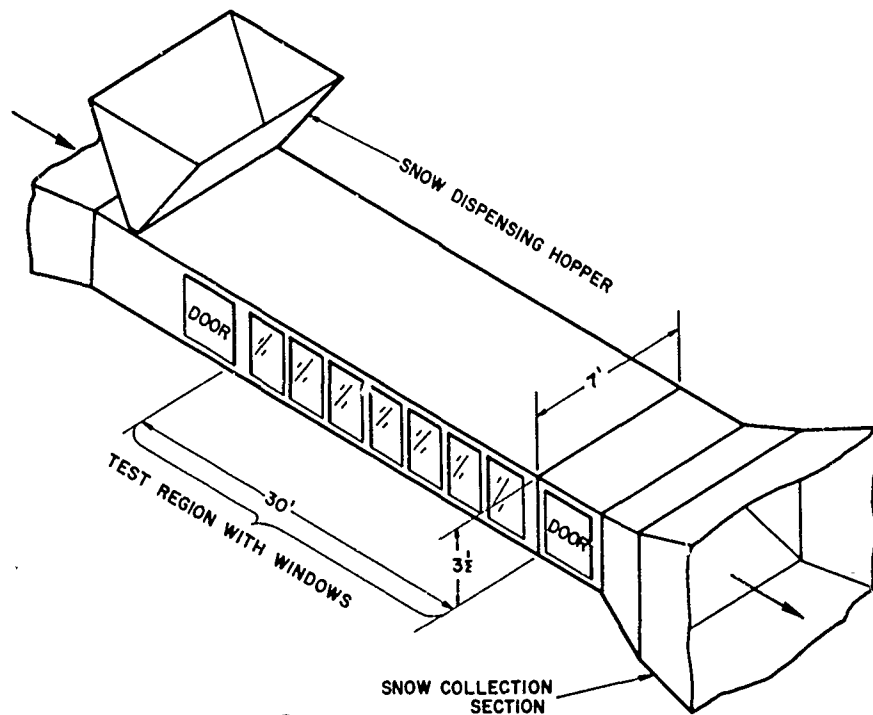


Figure 31. Sketch of the test section of the $3\frac{1}{2} \times 7$ ft wind tunnel used by Strom et al.⁸⁷

series of tests (Keitz, 1960).^{42,43,44} The tests and the analyses on which they were based are discussed by Strom, Kelly, Keitz and Weiss⁸⁷ (1962). A further theoretical study of modeling was made in 1960 by Silberman.*

Strom et al.⁸⁷ developed their scale factors by dimensional analysis and by consideration of theoretical equations of motion for saltation. Applying dimensional analysis to the turbulent suspension process they obtained a scale factor relating particle path dimensions to dimensions of the test object (e.g., model building) as a function of the following dimensionless groups:

1. Ratio of particle diameter to linear dimension of test model.
2. A Froude number based on particle size and velocity.
3. Ratio of particle and air densities (discarded as a negligible factor).
4. Ratio of particle velocity to fluid velocity at the particle (implying a scaling of aerodynamic drag).
5. A Reynolds number based on particle size and velocity (again with drag implications).

More uncertainties entered the consideration for saltation, but the basic problem involved scaling of the velocity profile in a turbulent boundary layer on an aerodynamically rough surface. The elasticity of the particles was considered to be important, and under this assumption the scaling condition for preservation of the velocity field was equality of the coefficient of restitution between particle and surface for both prototype and model.

*Unpublished USA SIPRE manuscript "Modeling blowing snow," by E. Silberman.

In scaling surface creep of snow particles, the threshold friction velocity, at which surface shear is just sufficient to roll grains along, was considered using Bagnold's³ condition derived originally for sand grains. If the ratio of particle and air densities is the same for prototype and model, scaling is in accordance with a Froude number, assuming that the surface is "rough." The surface regime was discussed in terms of the roughness Reynolds number.

A second analysis of saltation scaling was made by considering the equations of motion in vertical and horizontal directions and imposing a requirement for similarity of trajectories. The results showed that, after a suitable geometric scale had been chosen, the ratio of model and prototype velocities was given by the square root of the ratio of linear model dimensions. A similar condition defined the scale for boundary layer thickness and also gave a relation between particle sizes and densities for model and prototype. The final requirement was for equality of coefficient of restitution in model and prototype.

The test program was finally set up with a linear scale of $1/10$, so that the required velocity scale given by the Froude number was $\sqrt{1/10}$. A painstaking search was then conducted to find a suitable material for model snow, potential materials being tested for size, density, free fall velocity, and threshold velocity and coefficient of restitution. The best compromise found was powdered borax which, in the commercial form ($\text{Na}_2\text{B}_4\text{O}_7 \cdot 10\text{H}_2\text{O}$) available in bulk, had a mean particle size of about 200μ and a free fall velocity estimated as about 100 cm/sec . It now appears that the particle size was too large, but at the time of the work the size of typical blowing snow particles had been overestimated by an order of magnitude. It also appears that, even if the elastic collision theory is valid, the role of saltation was somewhat exaggerated. The models tested included barracks buildings on piles, an elevated radar station, fuel tanks, and snow fences, all in various groupings and orientations.

Odar^{67,68} has reviewed the general scaling requirements for the simulation of drifting snow together with the similitude requirements for certain specific characteristics. His review, which reiterates some of the foregoing points, is summarized below.

Geometric similarity calls for a constant ratio between corresponding linear dimensions \underline{L} for prototype and model.* Thus,

$$\frac{L}{L_m} = \text{constant.} \quad (21)$$

A problem immediately arises, for geometric scales which permit models of large buildings to be placed in normal wind tunnels become unreasonable if applied to the scaling of blown snow particles, which are typically about 100μ in diameter, full scale.

Kinematic similarity requires a constant ratio between velocities \underline{V} which correspond in direction and position in the geometrically similar systems of model and prototype:

$$\frac{V}{V_m} = \frac{V_p}{V_{pm}} = \text{constant.} \quad (22)$$

This condition necessitates a scaling of the boundary layer and of the turbulent structure. If saltation is to be simulated according to current views (see p. 6), it implies that the coefficient of restitution must be the same for the model and the prototype.

*Symbols without subscript refer to prototype, those with subscript \underline{m} refer to model. The subscript \underline{p} denotes a particle dimension.

Dynamic similarity is obtained when there is a constant ratio between forces $\frac{F}{F}$ which correspond in direction and position in the geometrically and kinematically similar systems of model and prototype:

$$\frac{F}{F_m} = \frac{F_p}{F_{pm}} = \text{constant.} \quad (23)$$

Fluid stresses σ must be similar and, since force equals the product mass x acceleration, there is a consequent requirement for a relationship of fluid density ρ and velocity V :

$$\frac{\sigma}{\sigma_m} = \frac{\rho V^2}{\rho_m V_m^2} = \text{constant.} \quad (24)$$

Odar proceeds to a consideration of the threshold characteristics, which determine the stage at which particles will begin to be moved from a cohesionless surface by the wind. The independent variables taken are particle diameter d (spherical particles are assumed), air density ρ , particle density ρ_p , and g (gravitational acceleration). The dependent variable is critical shear stress τ_c , i. e., the surface shear stress which is just sufficient to start cohesionless particles moving. Dimensional analysis leads to the relation

$$\tau_c = (\rho_p - \rho) g d f(\rho_p / \rho) \quad (25)$$

which differs from Bagnold's expression only by the form of $f(\rho_p / \rho)$, which Bagnold regarded as a constant for his air and sand system. The value of $f(\rho_p / \rho)$ for a mixture of air and snow particles has not been determined.

Treating the surface roughness of the snow before loose particles begin to move, the sub-critical surface shear stress τ_0 is influenced by a parameter λ which is a function of the effective roughness:

$$\tau_0 = \lambda \rho V^2. \quad (26)$$

Applying the dynamic similarity concept to τ_0 and τ_c a criterion for the scaling of threshold characteristics is obtained:

$$\frac{\lambda \rho V^2}{\lambda_m \rho_m V_m^2} = \frac{(\rho_p - \rho) g d f(\rho_p / \rho)}{(\rho_{pm} - \rho_m) g_m d_m f(\rho_{pm} / \rho_m)}. \quad (27)$$

Equation 27 is valid only for turbulent flow over an aerodynamically rough surface; if the surface is smooth and a laminar sub-layer exists, air viscosity μ enters the analysis as a significant additional variable. A similar dimensional analysis leads to the condition

$$\frac{\lambda \rho V^2}{\lambda_m \rho_m V_m^2} = \frac{(\rho_p - \rho) g d \phi \left(\frac{\rho_p}{\rho}, \sqrt{\frac{\tau_c}{\rho}} \frac{d \rho}{\mu} \right)}{(\rho_{pm} - \rho_m) g_m d_m \phi \left(\frac{\rho_{pm}}{\rho_m}, \sqrt{\frac{\tau_{cm}}{\rho_m}} \frac{d_m \rho_m}{\mu_m} \right)} \quad (28)$$

where λ is now dependent on Reynolds number as well as effective roughness. The function ϕ remains to be determined for a snow-air combination.

Odar further considers similitude of streamlines and the duplication of snow concentration along them. Similarity of snow particle concentration \underline{n} is given by

$$\frac{\underline{n}}{n_m} = \frac{N \rho_p g d^3}{N_m \rho_{pm} g_m d_m^3} \quad (29)$$

where N is the number of particles per unit volume. For duplication of streamlines, conditions based on the vertical transport equation (see eq 9 and 10) must be satisfied; in the symbols defined on p. 7:

$$\frac{\epsilon_z}{\epsilon_{zm}} = \frac{v_z}{v_m z_m} \quad (30)$$

Rate of accumulation or scour can be scaled from the simplified vertical transport equation for net transfer per unit area (\underline{Q}) at the surface:

$$v_n - \epsilon_z \frac{\partial n}{\partial z} = -Q \quad (31)$$

The scale factor is

$$\frac{Q}{Q_m} = \frac{v_n}{v_m n_m} \quad (32)$$

Another scaling condition for accumulation rate \underline{Q} may be based on the rate of change of elevation of the snow surface \underline{h} and the bulk density of the deposited snow γ :

$$\frac{Q}{Q_m} = \frac{\gamma \dot{h}}{\gamma_m \dot{h}_m} \quad (33)$$

Odar also discusses time scaling in order to account for the densification of deposited snow which occurs in nature. The expression he adopts to relate bulk density with depth below the surface is unduly complicated; a linear relation may be substituted instead.

Duplication of particle paths in accordance with the equation of motion is offered as an alternative to streamline duplication from the vertical transport equation. An equation of motion for spherical particles in a turbulent fluid with large eddies and low frequency fluctuations is given following Odar and Hamilton⁶⁹. It takes account of gravity and fluid forces, including those due to velocity and acceleration of a particle relative to the fluid. The scale factors derived from the equation of motion correspond to two of those derived by Strom et al. from their dimensional analysis for turbulent diffusion:

$$\frac{V^2}{L g} = \frac{V_m^2}{L_m g_m} \quad (34)$$

$$\frac{C_D \rho V^2}{d g \rho_p} = \frac{C_{Dm} \rho_m V_m^2}{d_m g_m \rho_{pm}} \quad (35)$$

where C_D is a drag coefficient.

The work of Strom et al. and Odar has provided a sound basis for the design of model tests, although some experimental work must still be carried out to determine unknown functions. It seems unlikely that all of the scaling criteria can be satisfied simultaneously, so that compromises will be necessary for tests in the immediate future. There appears to be a current need for more field data on which to base model tests.

Snow fences and their deployment

Snow fences fall into three main classes:

1. Collectors, arranged normal to the direction of the snow-bearing wind so that snow will settle out in the eddies they create.
2. Blowers, constructed in such a way that they accelerate wind flow locally and thus keep the surface nearby scoured free of snow.
3. Deflectors, leaders or guide walls, aligned at angles to the wind, so that they deflect the flow from the area to be protected.

Collecting fences are made from a wide variety of materials in various parts of the world. Most of them are of open construction, so that the wind can blow through them while the general flow is displaced upward. The advantage of open construction is twofold: it turns out to be more efficient than a solid barrier, and it is economical in material. The percentage of the total presented area of a fence which is of solid material is referred to as the density of the fence; the common range of fence densities for highway protection is 40 to 60%, but 20 to 25% has been recommended for increasing the snow cover on agricultural land.⁴⁶ Fence heights commonly range from 1 to 2 meters (3 to 6 ft), the quantity of snow collected increasing with height. A gap of 15 to 40 cm (6 to 16 in.) between the ground and the bottom of the fence is usually found to be advantageous.

Figures 32 to 41 show some common types of construction for snow fences (numerous other variations are illustrated by Schneider⁸²). As long as the fence density is about 50% there is probably little to choose between them in terms of efficiency; the preferred construction will be that which is most economical or expedient. For some purposes, e.g. establishing optimum fence arrangements by trial, moveable fences have advantages. Although not illustrated, dense netting is also used; advantages include ease of installation and storage. Meshes of coconut fiber and wire netting woven with steel strip are manufactured as snow fencing in Europe⁸². Plowed swales of snow have been used to good effect as collector fences.

Most snow fence protection is provided as an afterthought, but it should now be possible to anticipate snowdrifting problems in the design stage of such works as road construction. For permanent protection, easy maintenance and good appearance there seem to be decided advantages in the use of trees and shrubbery as snow collectors, provided there is space available. It might be noted in passing that if shrubs are planted to protect an area they need not be in regimented arrays; discrete clumps may well be more efficient than a continuous hedge, and they are certainly more appealing aesthetically. The tree chosen for snowdrift control should be one suited to local soil and climatic conditions, and it should be a variety which retains a fairly high "density" during winter. Evergreens are obviously useful, but there are also deciduous trees suitable for drift protection. Finney²⁸ gives some guidance on the layout of snow hedges and shelter belts.

Collecting fences act most efficiently when at right angles to the wind, so that the chief aim in placing them is to achieve a normal arrangement on the upwind side of the road, which itself may be at any angle to the wind direction. Variations of wind direction have to be taken into account, as have local disturbances of wind direction caused by deflection around hills, embankments, woods, etc. Setting out a fence system calls for good judgement and local knowledge based on an understanding of the principles involved (a discussion of snow fence performance comes later). However, most arrays are variations and combinations of three basic arrangements, known as parallel, staggered, and wing fences (Fig. 42). Further details of fence arrangements which have proved valuable under special circumstances are given in the literature (Bekker,⁹ Pugh⁷⁵, Price⁷³, Schneider⁸²).

BLOWING SNOW

The principal type of blower fence is an inclined "table top" arranged so that the wind blowing beneath it is accelerated across the area to be protected. The "table top" is set on posts (Fig. 43) and inclined at 15° to 40° with the highest edge facing into the wind. The lee edge is usually about 1.2 m (4 ft) above ground. The blower is placed close to the protected area—about 3 to 6 m (10 to 20 ft) away on the upwind side—and it must be carefully oriented to the direction of the snow-bearing wind. It becomes ineffective if the wind blows from directions other than that for which it was set, so that it is only useful in particular circumstances. The most common applications for blowers are protection of shallow cuttings (Fig. 41) and accumulation control on avalanche paths. In Europe blowers are called "pupitres" or "Pultdächer" (desks, or desk roofs) after their resemblance to old style school desks.

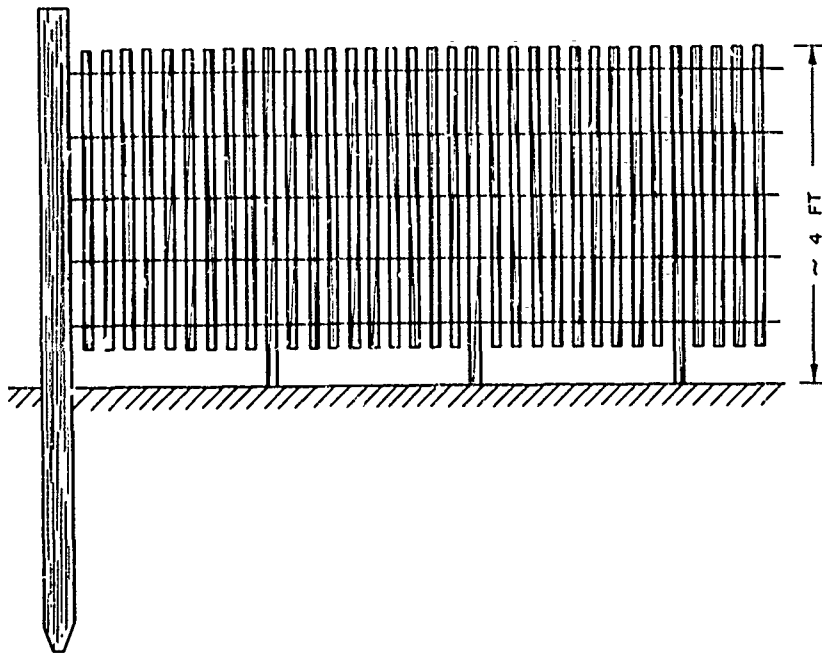


Figure 32. Vertical slat fence. This type of fence, consisting of wooden slats wired together, is common in the U. S. and Canada. It is light, cheap, and can be rolled up for storage in the summer.

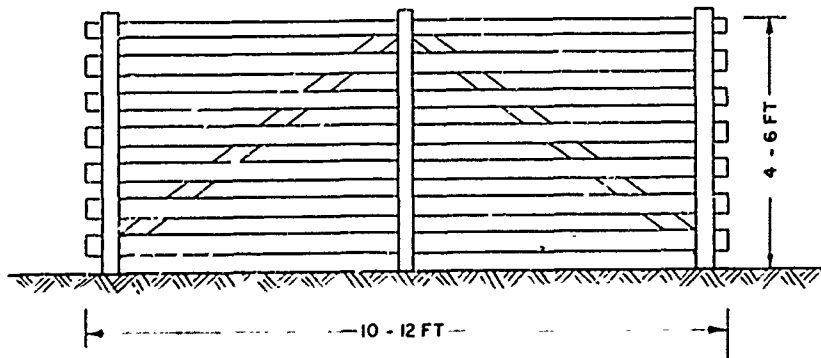
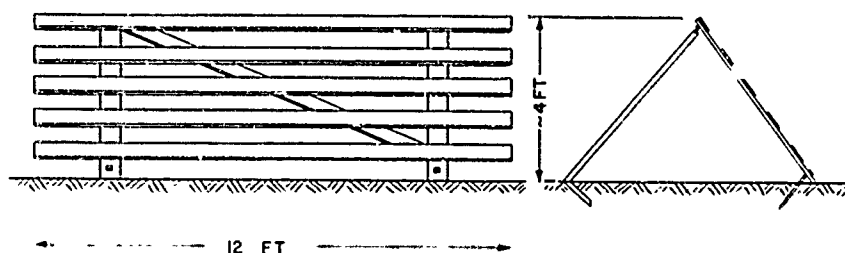
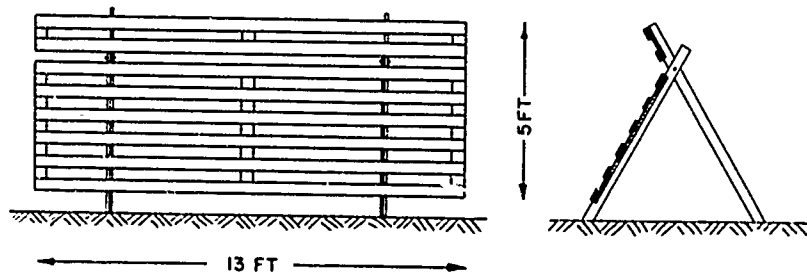


Figure 33. Horizontal slat fence of a type common in Scandinavia.

Deflector fences are not used very often, so that it is hard to generalize about their construction and layout. They must be solidly sheeted and smooth, as well as fairly high—say about 3 m (10 ft). Angles to the wind and curvature should be such that the wind flow is deflected smoothly. Deflectors must be used with great caution, for if a snow-bearing wind should come from a direction other than the design direction the deflector will become a collector, doing more harm than good. A similar effect might be produced by a "high" snow storm (appreciable snow concentration at a height of 10 m (33 ft) or more), irrespective of wind direction. Schneider⁸² shows how guide walls are employed on the Parsem railway in Switzerland, where they have proved valuable. Deflectors have been used with some success at ice cap sites to keep clear piles of supplies and entrances.



(a)



(b)

Figure 34. Inclined slat fences of the Swedish type.

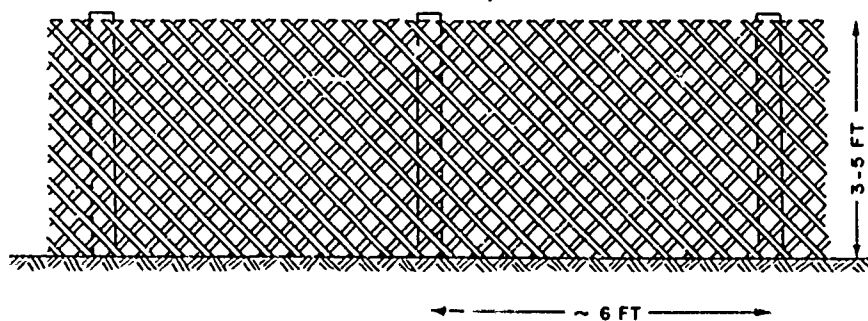


Figure 35. Trellis fence.

BLOWING SNOW

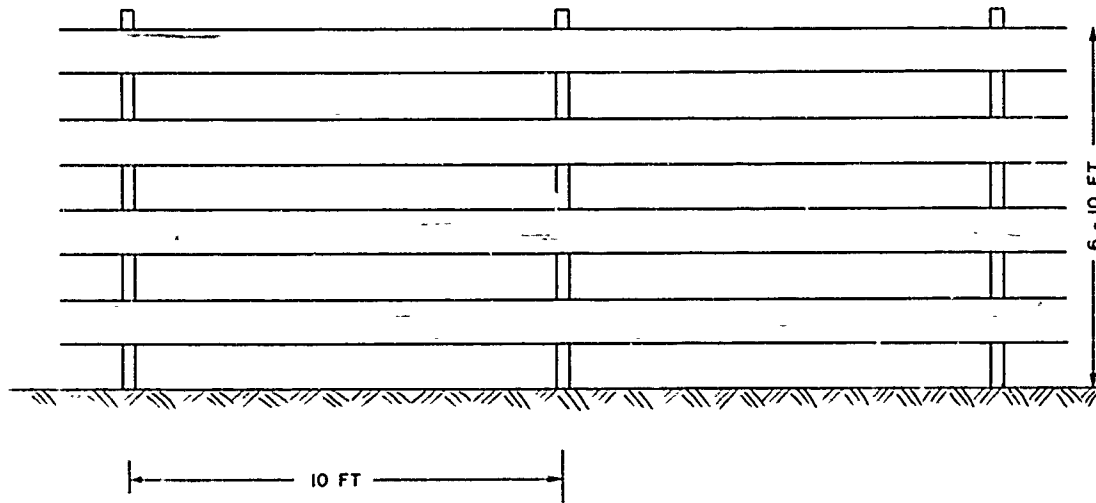


Figure 36. Fence made by stapling strips of strong paper onto wooden posts (the paper used in tests by USA SIPRE was heavy-duty packing paper consisting of sisal strands and bitumen sandwiched in brown paper).

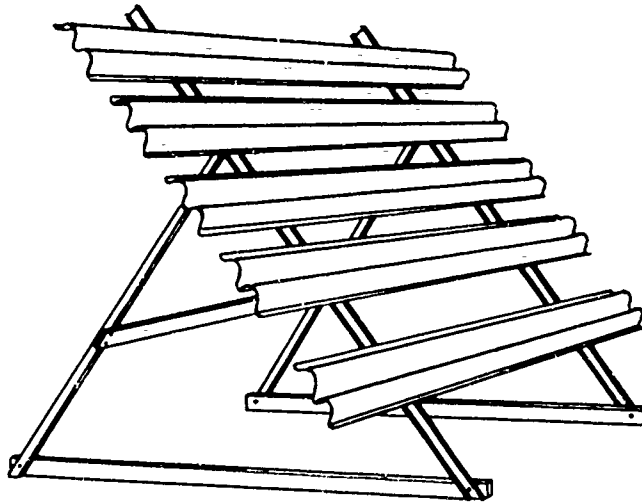


Figure 37. Movable section of corrugated metal fence used by USA CRREL for local protection in Greenland. Similar fencing, which is available commercially, is used for railroad protection in the U. S.

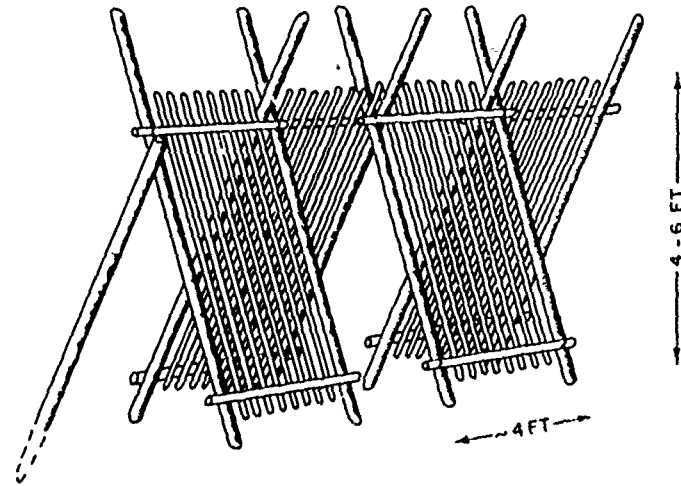


Figure 38. Old style Russian snow fence.

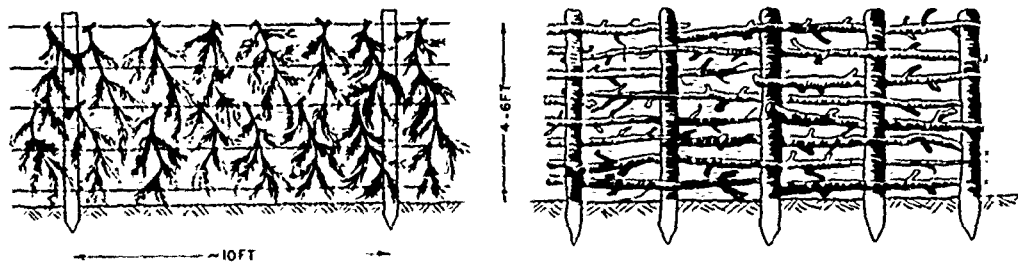


Figure 39. Pine bough fences.

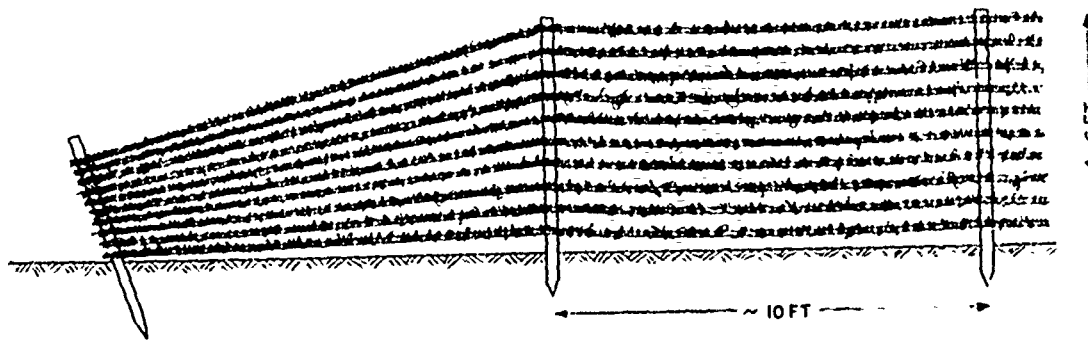


Figure 40. Straw rope fence.

BLOWING SNOW

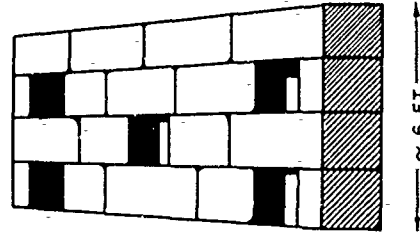


Figure 41. Snow block fence.

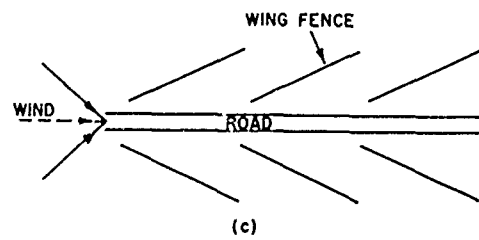
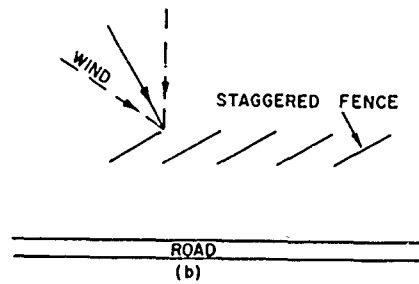
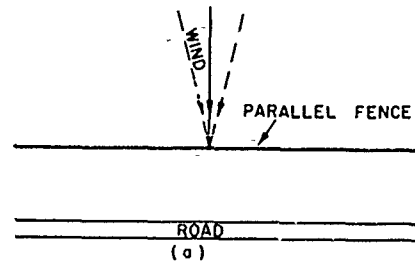


Figure 42. Basic arrangements of snow fences for protecting roads or railways. (After ref.76)

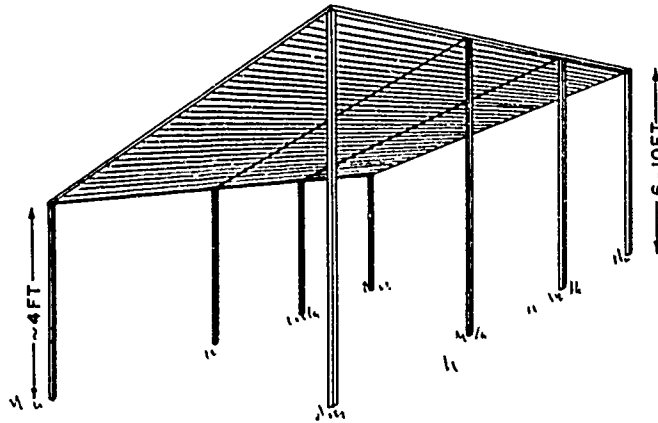


Figure 43. Blower fence ("Pulldäch" or "pupitre").

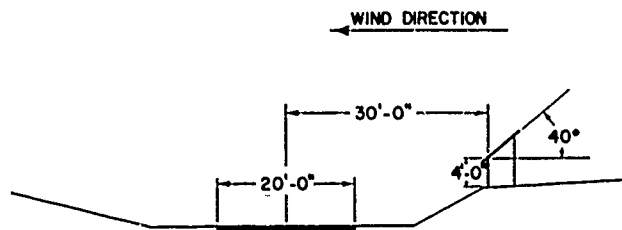


Figure 44. Example of blower fence protecting a road in a shallow cutting. (After ref 72).

Performance of snow fences

If snow fence protection is to be designed rationally it is necessary to have information on the size and shape of the drifts formed when wind conditions, fence type, and fence arrangements vary.

The shape of drifts formed by typical snow fences. Figure 45 illustrates the development of windward and lee drifts at a common type of Russian snow fence made from vertical wooden slats to a density of approximately 50%. The fence was set without a gap beneath it. Figure 46 gives some other examples of drifts formed by a solid fence and by vertical slat fences of various densities, all set originally with a 20-cm (8-in.) gap beneath them. The drifts continue to build up until they reach a form sufficiently streamlined to inhibit further deposition, at which stage the fence is said to be "saturated." There has been some discussion of the form of a longitudinal drift profile at saturation, ichthyoid curves and mathematical rose functions being offered as representations^{74,75}.

It has been reported⁹⁷ that windward drifts are "harder" than lee drifts, presumably because density is higher in the windward drift.

BLOWING SNOW

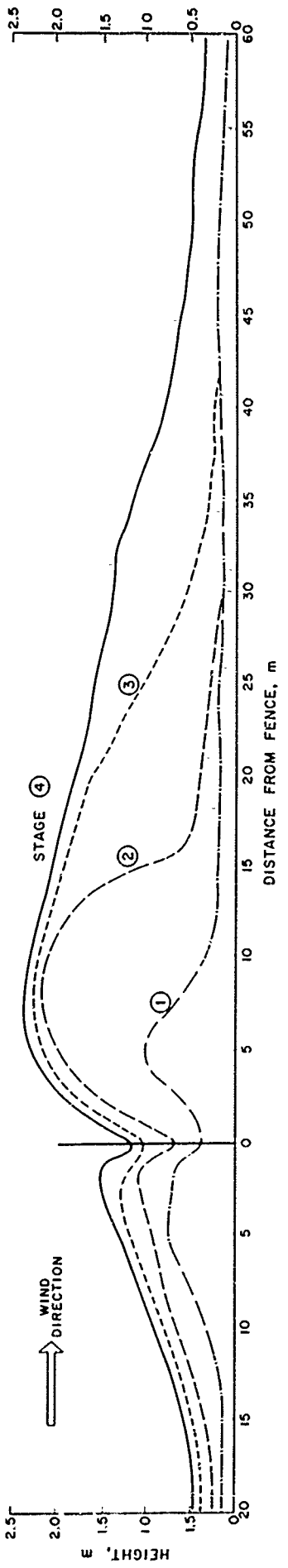


Figure 45. Snowdrift generated by a snow fence of standard Russian construction (vertical slats, approximately 50% density, no gap beneath the fence). (After ref #6).

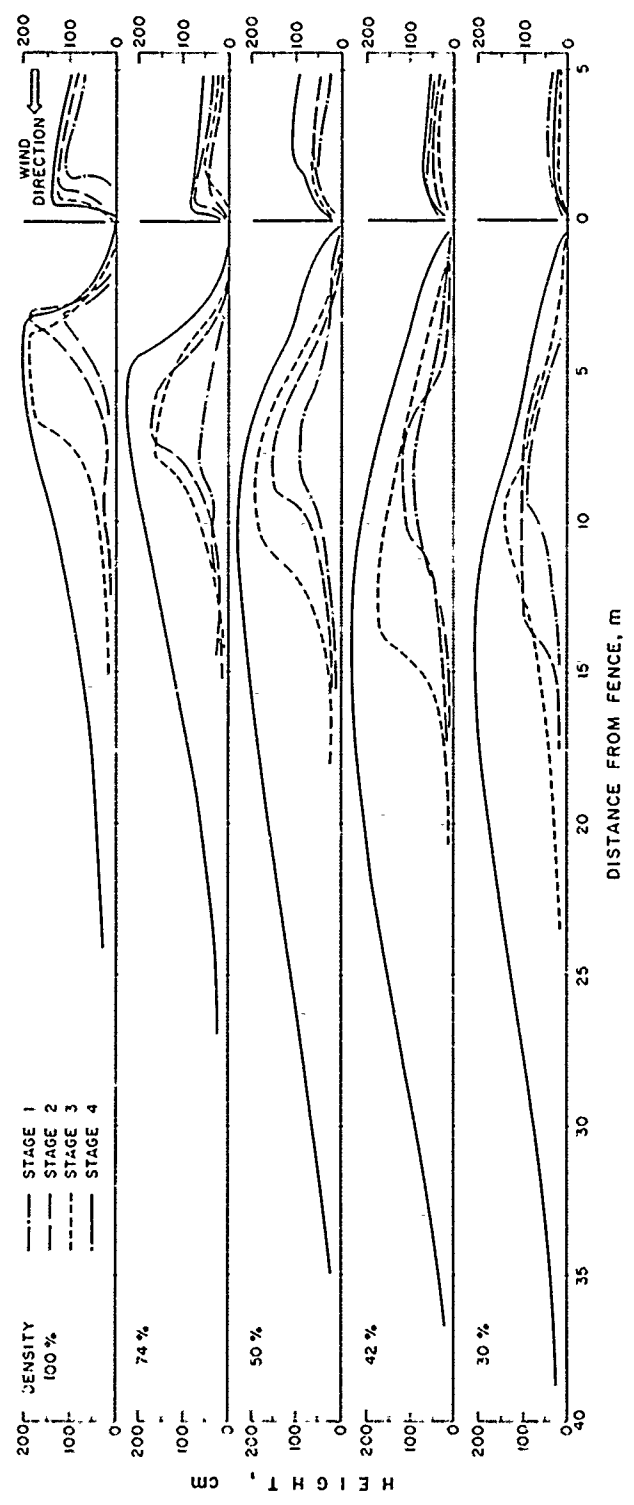


Figure 46. Snowdrifts generated by solid snow fences and by vertical slat fences of various densities (all 1.87 m high with a gap beneath them of 20 cm ± 5 cm). (From Price 74).

Effect of wind velocity on the performance of fences. Information from the literature is inconclusive on this point. From model tests in the wind tunnel Finney²⁷ decided that the sectional area of the eddy zone behind a fence is unaffected by wind speed over the range which is of interest, but the issue seems to be clouded by his additional finding that the deepest part of the lee drift moves further from the fence as wind speed increases. Also from model tests, Nøkkentved⁶⁶ found some shortening of the drift as wind speed increased. Kungurtsev⁴⁹ claimed evidence of an increase of drift length with wind speed, and made the somewhat dubious supposition of a simple linear relation between drift length and velocity. This last relation was said to be in agreement with wind tunnel results obtained in 1933/34 by Moroshkin⁶ (reference listed, but not available to this author). From general hydrodynamic considerations there can be little question that the size and structure of the eddy zone do change with velocity and acceleration of the wind, but whether the effects on snowdrifts are significant over the normal range of wind speeds remains to be shown. Practical men seem to feel intuitively that drifts will be longer, and perhaps relatively shallower, at high wind speeds.

Effect of fence density on snowdrifts. In allowing air to pass through it, a fence of open construction creates numerous vortices additional to those produced by the general boundary layer separation. The eddy zone is longer and shallower than that formed behind a solid fence of the same height (Fig. 26) and, unless the fence density is very low, more snow is deposited on the downwind side. Figure 46 shows profiles of the drifts formed by fences of various densities.

In Figure 47 available data^{5,74} have been used to relate the lengths of lee drifts non-dimensionally with fence density (the implicit assumption of geometric similarity is felt to be justified for a limited height range). For fence densities greater than 30%, both sets of data show an increase of drift length with decreasing density which can be represented approximately by

$$\frac{L_1 - L_2}{D_2 - D_1} = 0.16 \quad (36)$$

where L_1 and L_2 are relative drift lengths at fence densities D_1 and D_2 expressed as percentages. More specific relations given by the original authors may be misleading, since they disagree quantitatively.

While long shallow drifts are desirable for some purposes (e.g., covering farm land with snow) the usual requirement is for deposition of maximum quantity of snow, and the volume of a drift must obviously fall off as fence density tends to zero. The variation of collecting efficiency with fence density is investigated by considering the area of the longitudinal profile of a drift (which is equivalent to volume per unit length of fence). In Figure 48 the data of Croce^{15,16} and Price⁷⁴ have been used to prepare a dimensionless plot showing variation of lee drift area with fence density. Drift area for each density has been normalized with respect to the corresponding drift area for a density of 40%, which happened to be a common density for two of the studies. Price's data indicate an optimum density of about 40%, confirming an earlier report by him⁷³. Croce concluded from his scattered 1942 data that 50% was optimum. His more consistent data for 1943/44¹⁶ indicate an optimum density between 50% and 60%, which is in agreement with the findings of Hallberg³⁷ and Nøkkentved⁶⁶.

While the lee drift of a snow fence is usually regarded as the significant one, the contribution of the windward drift must also be considered. Croce¹⁵ plots the area of windward drifts against density, but there is hardly a significant variation. Price plots against density the ratio of windward drift to leeward drift, showing an increase in this ratio by roughly an order of magnitude as density increases from 30% to 100%. In the early stages of drift development the windward drift for solid fences was bigger than the lee drift, but at saturation it was only about one-third

BLOWING SNOW

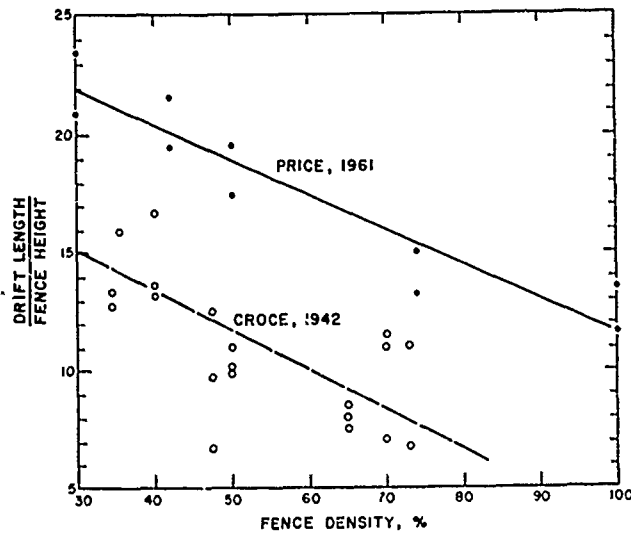


Figure 47. Ratio drift length/fence height plotted against fence density. (Data from ref 15, 74, 82).

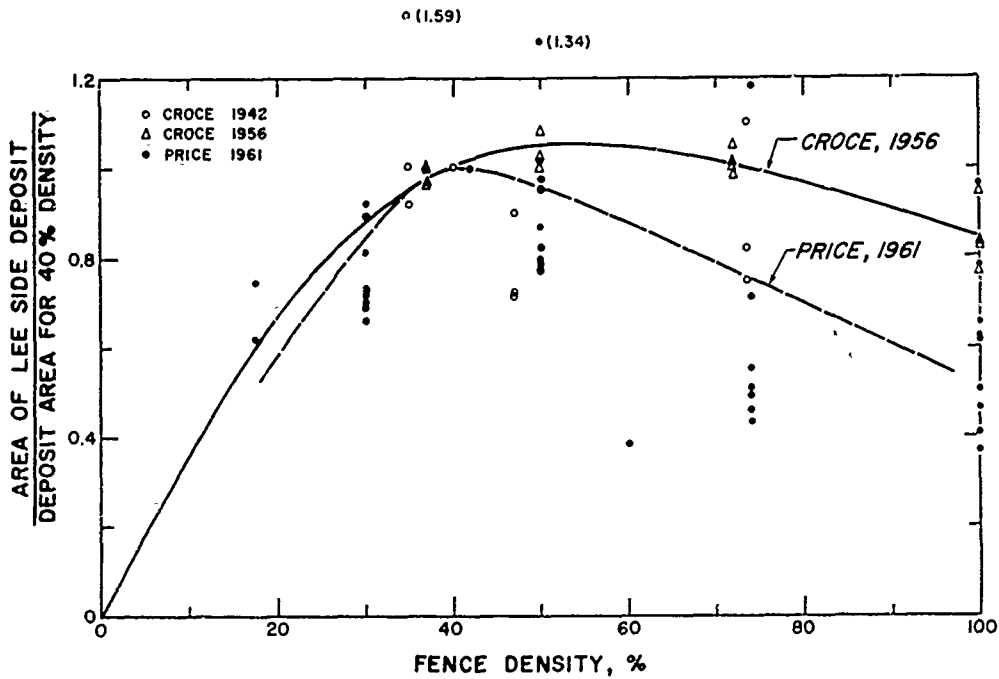


Figure 48. Relative area of lee drift plotted against fence density. (Data from ref 15, 74, 82).

the size of the lee drift (windward drifts for 30% and 40% density fences were insignificant compared with the lee drifts). This increase in the size of the windward drift relative to the lee drift with increasing fence density reduces the apparent advantage of the lower density fences illustrated in Figure 48. It seems from the limited available data that there may be only a marginal increase in the total amount of snow collected when fence density is reduced, although the pattern of deposition is affected appreciably.

A significant advantage of lower density fences is economy in materials. Another benefit obtained when a medium-to-low density fence is set with a gap beneath it is freedom from clogging, i. e., snow does not collect directly against the fence, so that the fence can be removed and relocated if necessary. In Russia, where fences are used to provide a distributed snow cover on fields, fence densities of 25% have been found to be desirable in multi-row arrangements^{46,47}. The low density fences, which are set out downwind of a row of standard fencing, produce long shallow deposits. Other fences have been built in Russia with 50% density on the upper half and 25% density on the lower half,²⁵ the purpose being to save material and prevent clogging, as mentioned above.

Influence of gaps beneath fences. It is well known that a gap between the ground and the bottom of a snow fence increases the working life of the fence by delaying saturation. It is also generally recognized that the lee drifts formed behind a fence with a gap are longer, shallower, and further downwind from the fence by comparison with those formed at fences set directly against the ground. It is common practice to set fences with a 15 to 20 cm (6 to 8 in.) gap beneath; adoption of this gap width probably owes much to Finney's²⁷ model tests which predicted maximum efficiency with a 15-cm (6-in.) gap, although it also appears to be a convenient height for uninformed fence users who are concerned chiefly with avoiding rotting of wet wood.

Croce¹⁶ made tests with 50% density fences set at 0, 20, and 40 cm (0, 8, 16 in.) above the ground; the drifts were displaced further away from the fence as the gap width increased, but there was no significant difference in the quantity of snow collected. Martinelli⁵⁶ placed 42% density fence with gaps of 30, 5, 61 and 122 cm (1, 2, 4 ft) beneath. The upwind end of the lee drift and the position of maximum snow depth both moved further from the fence as gap width increased. However, there was a very considerable decrease in the volume of the lee drift as gap width increased (Fig. 49). There was also a sharp decline in the maximum depth of the lee drift as gap width increased (Fig. 50). For most locations it is unlikely that there will be much advantage in setting fences with gaps greater than 30 cm (1 ft) or so, but in places subject to extreme frequency and duration of snowdrifting there may be value in large gaps, since the gap should eventually plug with snow and allow the fence to perform thereafter as a zero-gap fence.

Variation of drift size with fence height. The relation between drift length and fence height is an important one practically, since it is needed to determine the optimum position of a fence. Two or three rival formulas are available, but they appear to be based on remarkably little evidence. Figure 51 summarizes the published information for 50% density fences at or approaching saturation.

After adjusting field data for fences of various densities to give a plot for 50% density, Croce fitted a straight line and suggested

$$L = 11 + 5h \quad (37)$$

where L and h are length of lee drift and fence height in meters respectively.

Price found from his field data a similar type of relation

$$L = 22.1 + 6.5h \quad (38)$$

again in meters.

From wind tunnel tests both Finney and Nøkkentved obtained a simple linear relation

$$L = K_1 h. \quad (39)$$

In Finney's case K ranged from about 13 to 16.5, while Nøkkentved found it to be 25.

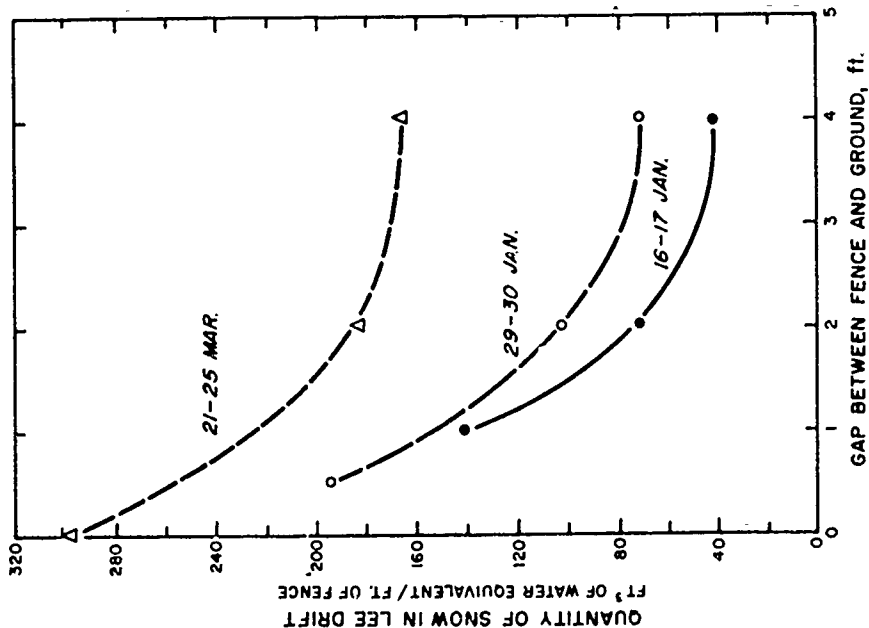


Figure 49. Quantity of snow collected in the lee of 6 ft of 42% density slat fencing set with initial base gaps of 1, 2, and 4 ft. Values for 6 in. and 0 gaps were obtained as the 1-ft gap gradually choked. Points averaged from two values at each of four locations. (From Martinelli⁵⁶).

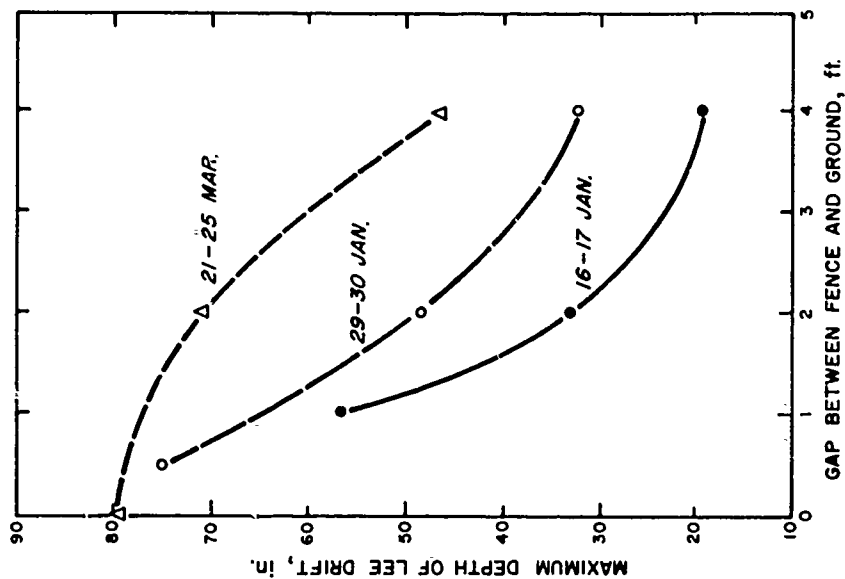


Figure 50. Maximum snow depth in the lee of 6 ft of 42% density slat fencing set with initial base gaps of 1, 2, and 4 ft. Values for 6 in. and 0 gaps were obtained as the 1-ft gap gradually choked. Points averaged from two values at each of four locations. (From Martinelli⁵⁶).

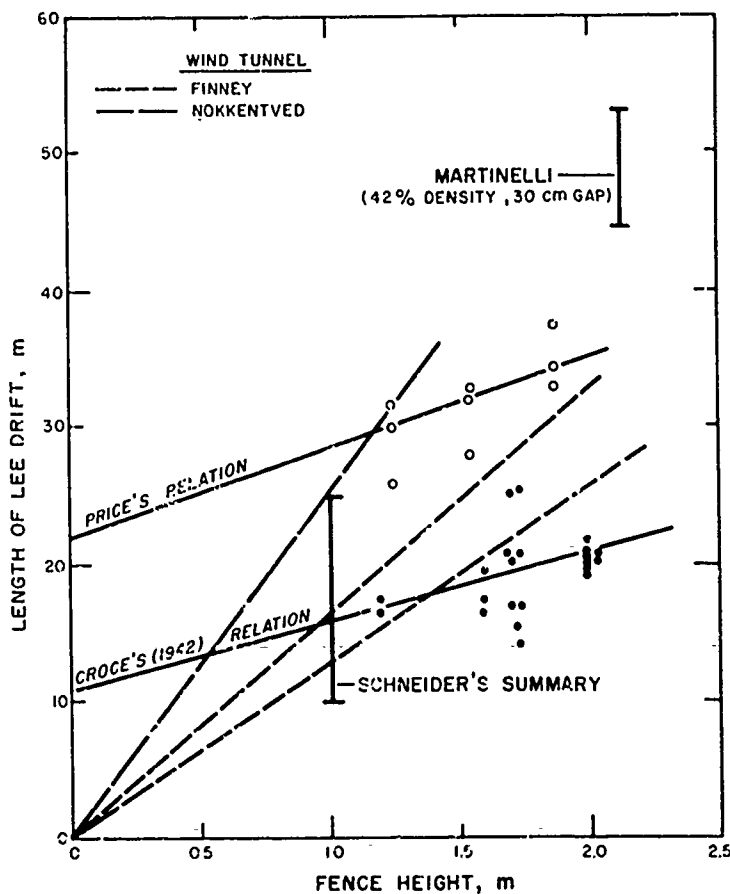


Figure 51. Distance from fence to end of lee drift plotted against total fence height for approximately 50% density fences approaching saturation. (Data from ref 15, 27, 56, 66, 74, 82).

Since they disagree quantitatively there seems to be no particular merit to eq 37 and 38; inadequacy of form can be seen by substituting $h=0$. As a working rule of thumb eq 39 might be adopted for the time being. Nøkkentved's expression $L=25h$ probably sets an upper limit, while $L=16h$, after Finney, is a compromise for the field data overall.

The range of h in a previous test has been too small to do much more than define a tangent to the curve for L versus h . Further experiments are required.

Another requirement is for an expression relating fence height and volume of collected snow, as given by the area for a drift profile at right angles to the fence line. Nøkkentved suggested that the drift area at saturation A might be given by

$$A = K_2 h^2. \tag{40}$$

This might be true if geometric similarity was maintained as fence height changed, but of course the structure of the boundary layer does not change. The limited findings of Price can almost be represented by a straight line,

$$A = 21h \tag{41}$$

where A is in square meters and h in meters. This may be compared with Croce's¹⁶ finding, viz.

$$A = 17.8h \tag{42}$$

in the same units. Here h is effective fence height, i. e., total height minus snow cover depth.

The maximum height of drifts on the lee side of slatted fences ($\approx 50\%$ density) exceeds the height of the fence itself. Price found the maximum height of saturated lee drifts to be 1.14 times the fence height. Russian observations give maximum drift heights up to 1.6 times the fence height (Kungurtsev,⁴⁹ after Dolgav).

Effect of fence inclination. From tests by Hallberg³⁷ it appears that inclining a fence away from the wind by a moderate angle produces a longer and lower eddy zone. Finney's²⁷ model tests gave 15° from the vertical as the angle for maximum drift length. Pugh⁷⁵ believes that fences inclined away from the wind at angles up to 30° from the vertical are acceptable; they produce longer and lower drifts, but the total catch is not significantly different from that for the same fence set vertically. If the angle of inclination exceeds 30° from the vertical, efficiency decreases. When fences are inclined towards the wind they tend to clog as snow builds directly against them on the windward side. When a fence with ("thick") horizontal slats is inclined, there is an increase of effective fence density. Schneider⁸² reports that German investigators consider the inclination of a fence wasteful.

Fence components.* Price⁷⁴ compared vertical slat fences of 42% density in which the slat width varied from 2.5 to 60 cm (1 in. to 2 ft). Changing the slat width from 2.5 to 22.5 cm (1 to 9 in.) had no effect on the size or shape of the drift, but with 30-cm (1-ft) slats there were corrugations in the drift as a result of a "shadow effect" behind the slats. This was more marked with slats 60 cm wide, and the total volume of the drift was reduced. Croce¹⁶ found no significant difference between slats of 3.0, 4.6 and 9.4 cm width in a 50% density fence, but wider slats were detrimental. Nøkkentved found no difference in performance between flat and rounded slats, but Pugh feels that rough-surfaced materials such as straw bundles or wool are more effective than smooth slats. Croce notes that natural hedges form bigger drifts than do fences of comparable height and density. It would be interesting to extend the study of fence materials to a smaller size range, perhaps testing dense meshes down to the size of mosquito netting (where clogging might become a problem). Croce¹⁵ felt that the finest possible structure should be most effective.

Multiple fences and movable fences. Fences arranged in two or more parallel rows are considered expedient both for spreading snow over an area and for combating adverse drifts in conditions which would require a single row fence to be very high^{46, 47}. The first row of fencing is of standard construction and about 50% density; succeeding rows have only 25% density. Heights range from 1 to 3 m (3 to 10 ft), and the spacing between fence rows is 30 times the fence height.

An alternative scheme for maintaining fence effectiveness in areas of heavy drifting involves the use of movable fence sections, such as the USA CRREL experimental fence shown in Figure 37. When the fence nears saturation it is extricated and set up again on top of its own drift, as illustrated in Figure 52. This procedure builds up a large concentrated drift. "Snow ramparts" have been built to heights of 10 m (33 ft) alongside Russian railways by successive erection of fences on drift crests (Chervinsky⁸²).

Drifting on highways and profile control

Most of the troublesome snowdrifts on highways are caused by eddy-forming obstructions alongside. Examples of such obstructions are fences and hedges close to the road, steep cuttings, buildings, snow banks left by plows, and features of the terrain which deflect and disturb the wind flow. The first step in alleviating drift problems involves removal of as many of these obstacles as possible, so that air flow across the pavement is unimpeded. When the margins of the highway have been cleared and smoothed as far as is practicable, further improvement of the situation by use of snow fences can be considered.

*Perhaps the ample available data on drag coefficients could be utilized to gain some insight into the drift-generating potential of objects when shape and size vary. Wind resistance represents an energy loss from the flow, and if the wind is "saturated" with snow particles a loss of energy should require that particles be deposited. Power expended by the flow as wind resistance might be equated to the power requirement for suspending snow to give an estimate of deposition.

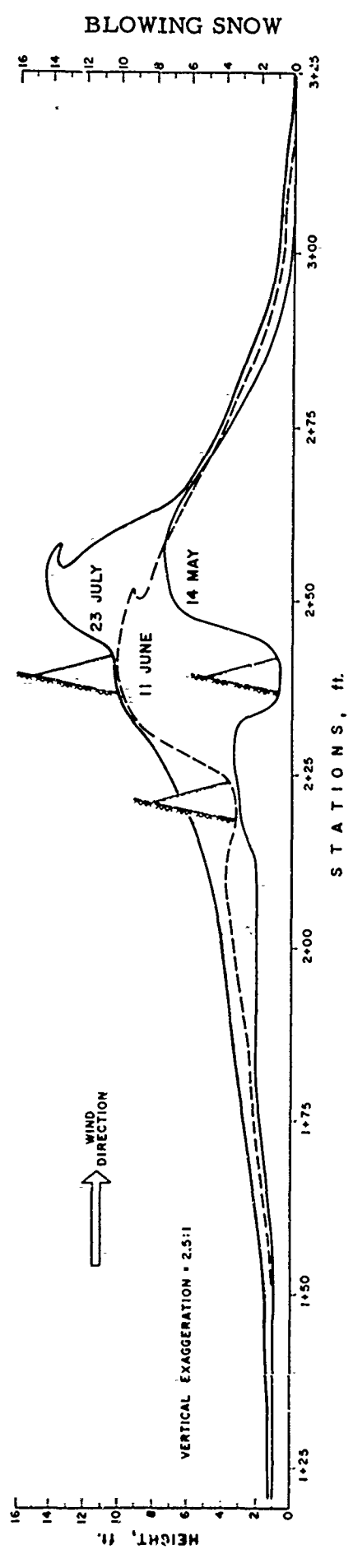


Figure 52. Large snowdrift generated in three stages by a movable fence of the type shown in Figure 37. (USA CRREL test on the Greenland Ice Cap.)
 (J. Hicks, USA CRREL internal lecture notes)

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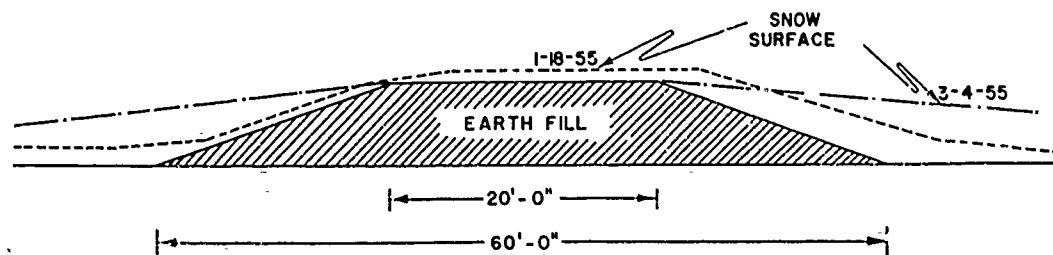


Figure 53. Drift formation against a test section of roadway elevated 6 ft above the surrounding country. (After ref 32).

In open country a road which is not elevated above its surroundings is difficult to keep free of snow; plowing provides only short-term relief, since the wind can soon refill the trough left by the plow. Swales of plowed snow subsequently aggravate the problem. A solution is to elevate the road. The desirable height of grade above mean ground level is determined from local conditions—principally the annual snow accumulation and local wind characteristics. In the parts of North America which experience a significant snow cover, efficient grade elevations are in the range 0.5 to 1.5 m (18 in. to 5 ft) (Gerdel³²).

If the side slopes of an elevated road are too steep, major eddies will form on both windward and leeward sides when wind blows across the road, and there will be some separation over the pavement itself. Finney²⁹ tested models of highway embankments in the wind tunnel to determine the effects of side slope variation. He found that 1:6 slopes created very little separation; with 1:4 slopes the flow disturbance was acceptably slight and the drifts formed were not severe. Rounding the lip and toe of the embankment improved the situation. Full scale tests were made in Michigan by the U. S. Army Snow, Ice and Permafrost Research Establishment (USA SIPRE)³² using sections of 1.8 m (6 ft) high road embankments which had side slopes of 1:2, 1:3 and 1:4. In all cases drifts accumulated against the embankments until the effective side slope was reduced to 1:9, after which there was no further drifting at the edges of the fill. Snow accumulation on the road surface reached a depth of 18 cm (7 in.), compared with 53 cm (21 in.) on the surrounding fields; it remained passable to wheeled vehicles without plowing (Fig. 53). A 1 m (3 ft) high embankment at the same site was not completely self-clearing, although it became passable late in the season while 69 cm (27 in.) of snow still lay on surrounding fields. It may be concluded from this study that, while side slopes tend to take care of themselves, the height of fill must be matched (equal to, or greater than) to the local accumulation for effective scouring.

In a cut section it is again necessary to consider the width of the cut and the angle of the side slopes. In wind tunnel tests Finney²⁹ found that wind flowing across a "step-down" bank formed an eddy zone which was always 6.5 times the height of the bank, irrespective of the wind speed or the slope angle (for slopes steeper than 1:6.5). The practical validity of Finney's finding has not been seriously challenged,* and recommendations for cut design (Fig. 54) have been based upon it^{73, 76}. Design considerations also include rounding of the lip and toe of slopes, and provision of storage space for snow which has to be plowed from the pavement. Deep cuts apparently enjoy a special immunity, and steep slopes are permissible⁸².

*Kungurtsev⁴⁹ claims that the length of the eddy zone increases with wind speed. In principle there is little argument that the shape of the eddy zone will vary with wind speed, but Kungurtsev gives no evidence to show that the effect is practically significant over the relevant range of wind speeds.

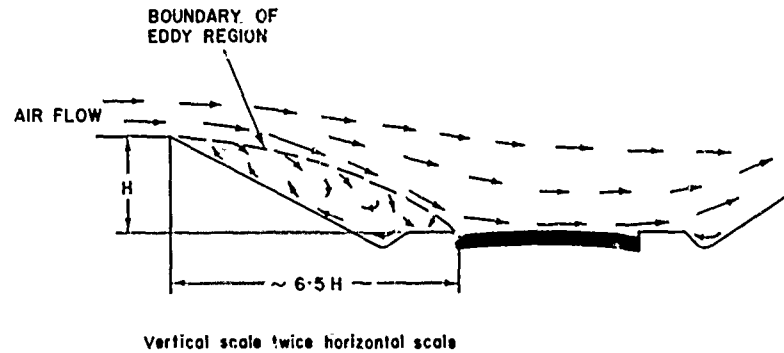


Figure 54. Drift-forming eddies in a cutting.
(Price⁷², following Finney²⁹).

Drifts generated by tall obstacles

The drift-producing characteristics of a tall obstacle can best be visualized by considering the flow disturbance in a horizontal plane, as opposed to the vertical plane considerations for "long" obstacles such as fences and embankments. On meeting a bluff obstacle the flow divides and forms eddies: on the windward side there are eddies immediately adjacent to the stagnation point, while on the lee side there is a wake of eddies. The broad features of the flow disturbance for cylinders and flat plates are illustrated in numerous descriptive texts on fluid flow, although usually for rather low Reynolds numbers. If the tall obstacle is angular in plan, there will be further separation downwind of the sharp corners. In the zones where forward velocity is translated to rotation, vertical exchange of momentum is destroyed and it becomes possible for snow particles to settle to the ground; this may occur to the sides of the obstacle as well as to windward and leeward. It often happens that the eddy backflow against parts of the obstacle's surface is too swift to permit deposition, so that a "moat" is left between the snowdrift and the obstacle in such places. Seen in plan, the outer limits of snowdrifts surrounding a tall obstacle define a streamlined shape enclosing the obstacle, so that the bulk of the deposited snow is in the "head" and the "tail" of the disturbance.

Drifting around buildings and similar obstacles

A long building lying normal to the wind direction behaves much like a steep embankment. Major eddies on windward and leeward sides form substantial drifts; providing there is no gap beneath the building the lower parts of drifts lie directly against the walls, although there may be "moats" between the building and the drift crests. There is some flow separation over the roof, but not much snow is likely to accumulate there. Eventually the building will become effectively streamlined by the drifts (like a saturated snow fence), and no further drifting will take place if the drifts are left undisturbed.

Most real buildings are not "long," however; their length and breadth are usually of comparable magnitude, so that "end effects" become significant. The three-dimensional flow disturbance is a result of vertical displacement of flow over the building and horizontal displacement around its ends; seen in plan the drifts formed will be similar to those occurring around "tall" obstacles, while in longitudinal section along the center line they will be similar to those occurring at "long" obstacles.

Drift patterns are further complicated when a principal axis of a building is not aligned with the wind, when successive storms come from different directions, and when buildings in groups interfere with each other.

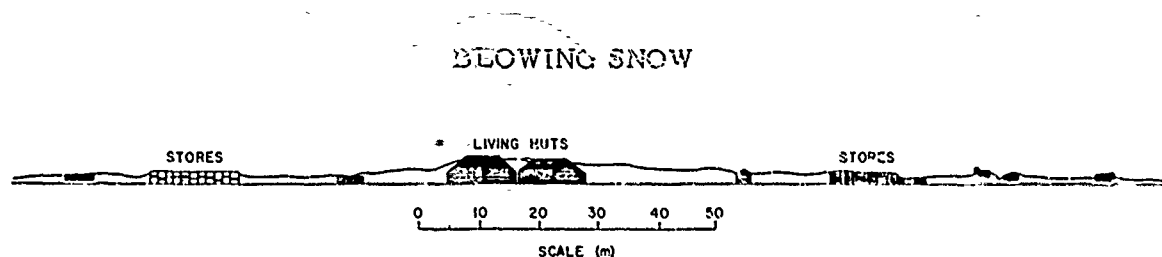


Figure 55. Cross-section of the compound drift which formed over Maudheim, Antarctica, in a 3-month period (section normal to the wind direction). (After ref 78).

As far as is known, drifting around real structures and groups of structures has not been studied systematically; one consequence of this is that the opinions and recommendations of different experienced observers, both published and unpublished, conflict to some extent. There is, for example, no general agreement on the proper orientation of a long structure; alignment both parallel and normal to the snow-bearing wind has been recommended, no doubt with justification for the particular combination of conditions envisaged but not explicitly stated. Rather than attempt another subjective appraisal, it seems preferable here to draw attention to some recorded observations on drift formation. References to model studies have already been given in the section on wind tunnel testing; in interpreting results of model tests, the limitations of existing modeling techniques should be kept in mind.

The drift problem may be very severe on polar snowfields, where there is continual accumulation and virtually no summer melting. Objects projecting from the surface can generate enormous drifts, so that surface structures are engulfed and, in some cases, crushed by the weight of snow. Although polar travelers have long been aware of the problem there have been few conscious efforts to describe it in detail, or to recommend solutions.

In 1955 Roots and Swithinbank⁷⁸ described the experiences of the Norwegian-British-Swedish Antarctic Expedition at Maudheim, a small base built on an Antarctic ice shelf. In setting out the huts and their surrounding supply dumps, snowdrifting was carefully considered in the light of both personal experience and the researches of Finney. Nevertheless, individual drifts coalesced and the base was buried in a great drift about 0.5 km wide and about 1 km long (Fig. 55). This result is almost inevitable with surface construction, but Roots and Swithinbank offer useful advice on ways to minimize the difficulties. Their points include: (i) separation of items which may be buried from those which must remain accessible, (ii) compact grouping of items which are allowed to drift over, (iii) spacing "accessible" items to avoid coalescence of drifts, (iv) elevation of objects to permit scour beneath them, (v) avoidance of unnecessary obstructions such as parked vehicles and spoil heaps from digging, (vi) estimation of storm directions from dunes as well as sastrugi (see "Deposition and Erosion"). Roots and Swithinbank also report that a line of boxes placed normal to the wind produced a much smaller volume of drifted snow than a similar line of boxes set parallel to the wind direction (Fig. 56).

In 1955 the U. S. Navy built four research stations on Antarctic snowfields. Prefabricated huts were erected on the snow surface in compact groups and, in anticipation of burial, they were connected by covered passages. It was believed that drifts would build up until the snow surface reached roof level and a streamlined drift profile was established; it was thought that thereafter further burial would proceed at a rate determined by the normal accumulation for the open snowfield. This was a reasonable assumption, but in practice repeated digging, dumping, and general disturbance of the surface in the camp area maintained the drift growth indefinitely. At Byrd Station excessive snow loads imposed by the great drift crushed the buildings, in spite of strenuous attempts to reinforce them. Figure 57 shows the drifting which occurred at Ellsworth Station in the first 9 months after construction: in the second picture the general level of the snow has risen by roughly 3 m (10 ft) in the camp area,

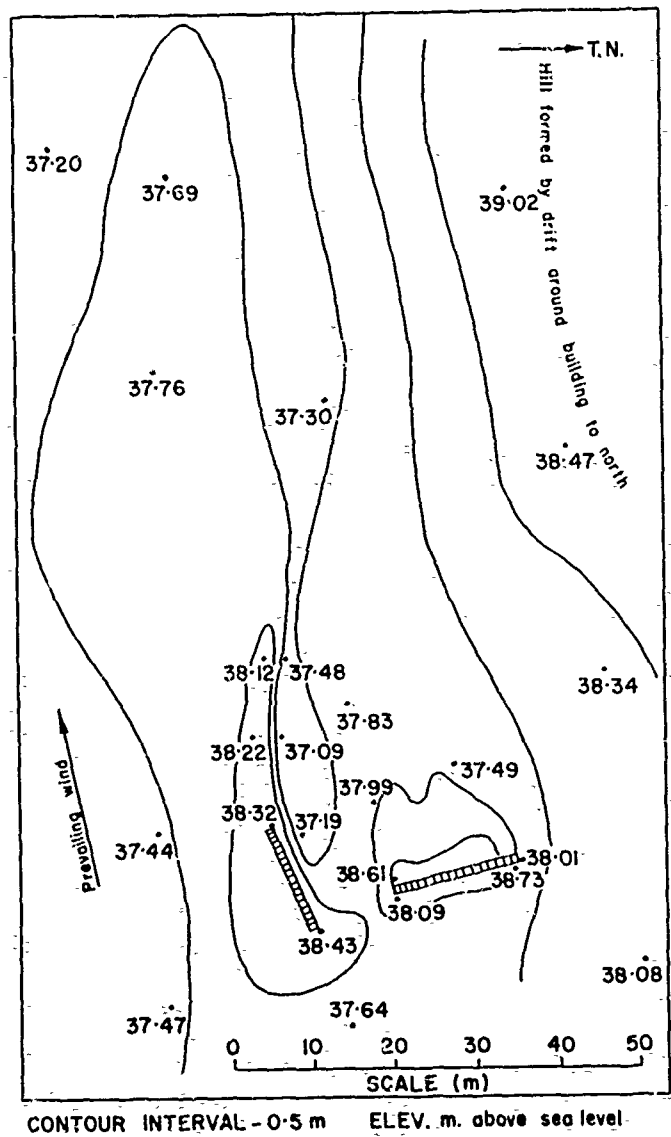


Figure 56. Contour plan comparing the drifts formed by long obstructions (lines of packing cases) normal and parallel to the wind direction. (After ref 78).

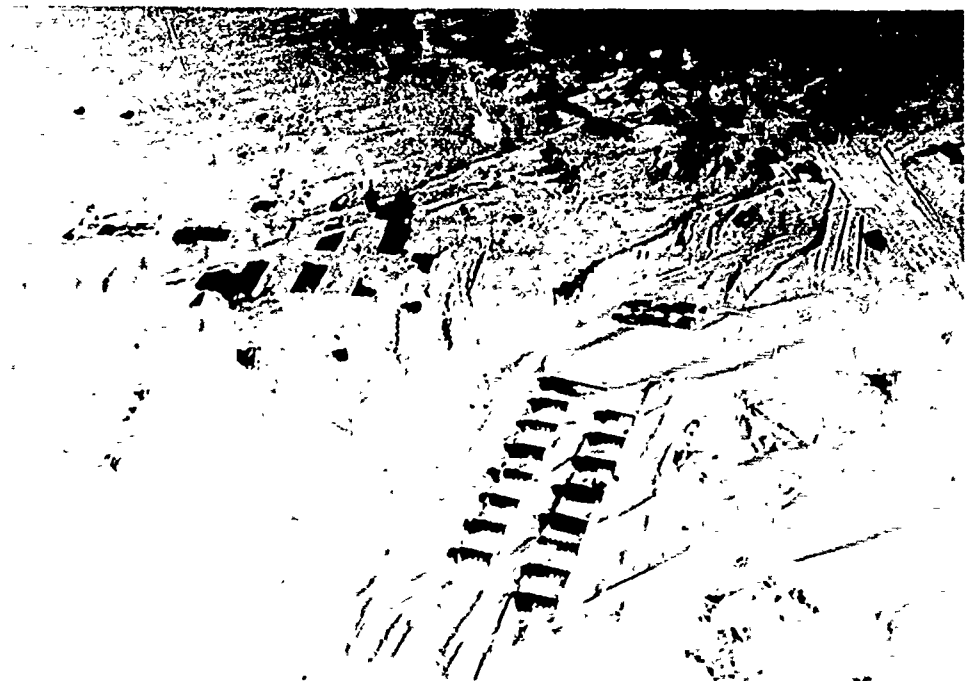
and although the roofs of buildings are flush with the surface, more drifts are being generated by new supplies, rubbish dumps, and particularly by spoil heaps thrown up by bulldozers. At these stations the rawin domes and the auroral observatories had to be kept above the surface in order to function. This requirement was fulfilled by setting the structures on tall columns, which could be extended when necessary to maintain the elevation above snow level. Figure 58 shows typical examples of these structures.

Although surface structures must inevitably create drifts, it is possible to operate a surface camp efficiently, at least for 2 or 3 years. This was impressively demonstrated at the New Byrd Station construction camp in Antarctica, where a combination of intelligent layout and immaculate maintenance relegated snow drifting to the role of minor nuisance. Sections of Jamesway building were joined to give a single structure of great length lying normal to the direction of storm winds (Fig. 59).

BLOWING SNOW

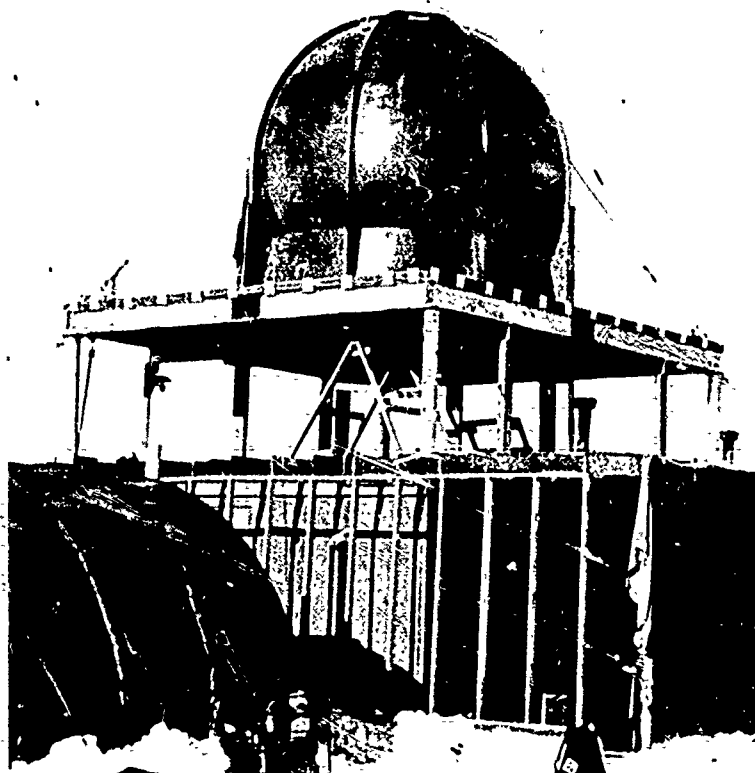


(a). The station under construction in February (late summer).



(b) The station in November (spring) from a different angle. In the 9-month period following construction the main buildings have been buried to the rooftops, and the construction camp of Jamesway huts is choked with snow in spite of clearing efforts. Bulldozing and resupply has created new obstructions and drift growth is continuing.

Figure 57. Drift formation at Ellsworth Station, Antarctica, in 1957.
(Official U. S. Navy photos).



(a) Rawin dome at the South Pole.



(b) Auroral observatory at Cape Hallett.

Figure 58. Small structures on extensible columns which can elevate and maintain them above the snow surface. (Official U. S. Navy photos).



Figure 59. Construction camp at the new Byrd Station, Antarctica. Jamesway sections have been connected to give a single long building at right angles to the wind direction. After rotary plowing the area has been kept perfectly smooth by continual grading.

Supply lines were set out downwind and parallel to the building with wide spaces between them. The supplies were elevated above the snow on long platforms formed by laying planks on empty fuel drums. The drift formed on the lee side of the building in winter was removed in spring by a rotary snow plow, and the entire camp area was graded to a perfectly smooth surface by weighted timbers dragged behind a tractor. Rigid discipline and regular maintenance ensured that there was no littering or disturbance of the surface, even while construction work went on at a rapid pace.

In recent years two major ice cap bases, Camp Century in Greenland and New Byrd Station in Antarctica, have been built in tunnels under the snow, following principles developed earlier on a smaller scale. This type of construction was adopted partly to avoid snowdrift problems and partly for other reasons. Potentially, undersnow construction is highly effective in eliminating snowdrifts, as has been found by a number of small expeditions in the past. In practice, however, some difficulties still remain. One of the chief problems lies in keeping access ramps free from snowdrifts, although efficient maintenance goes far to providing a solution. A more serious, but avoidable, problem arises from disturbance of the surface by supply dumps, vehicles, scrapped equipment, camp rubbish, and spoil heaps of snow.

When the DEWline was extended eastward from Cape Dyer in 1959-60, two stations on the Greenland Ice Cap were required. Since the huge scanning and communications antennas had to be kept clear of the snow, each of the stations was built above surface on columns. These gigantic structures, designated Dye 2 and Dye 3, each consisted of a single composite building with attached antennas elevated about 6 m (19 ft) above the snow surface on eight columns (Fig. 60). Jacks capable of lifting the buildings were incorporated into the columns, and there was provision for extension of the columns so that the design elevation (6 m-19 ft) could be maintained over a 10-year

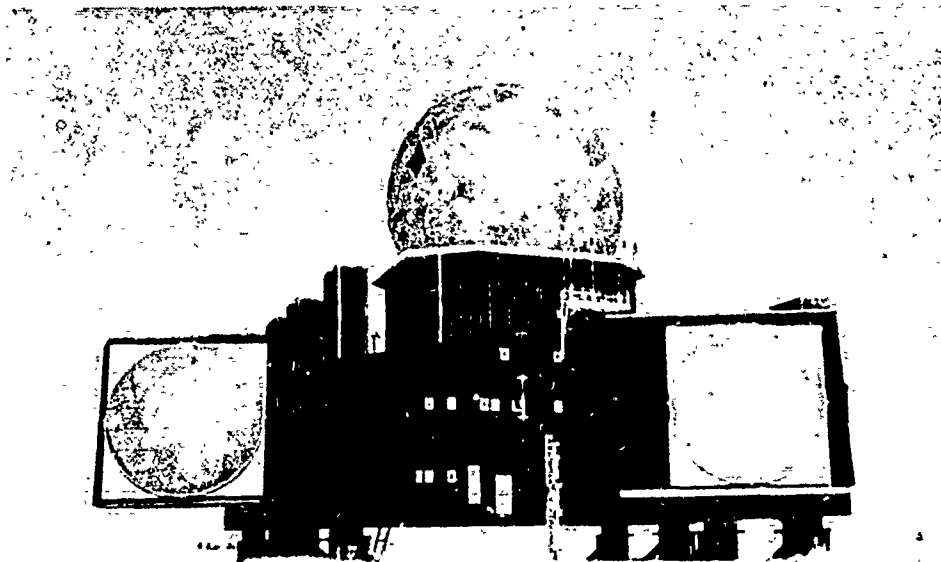
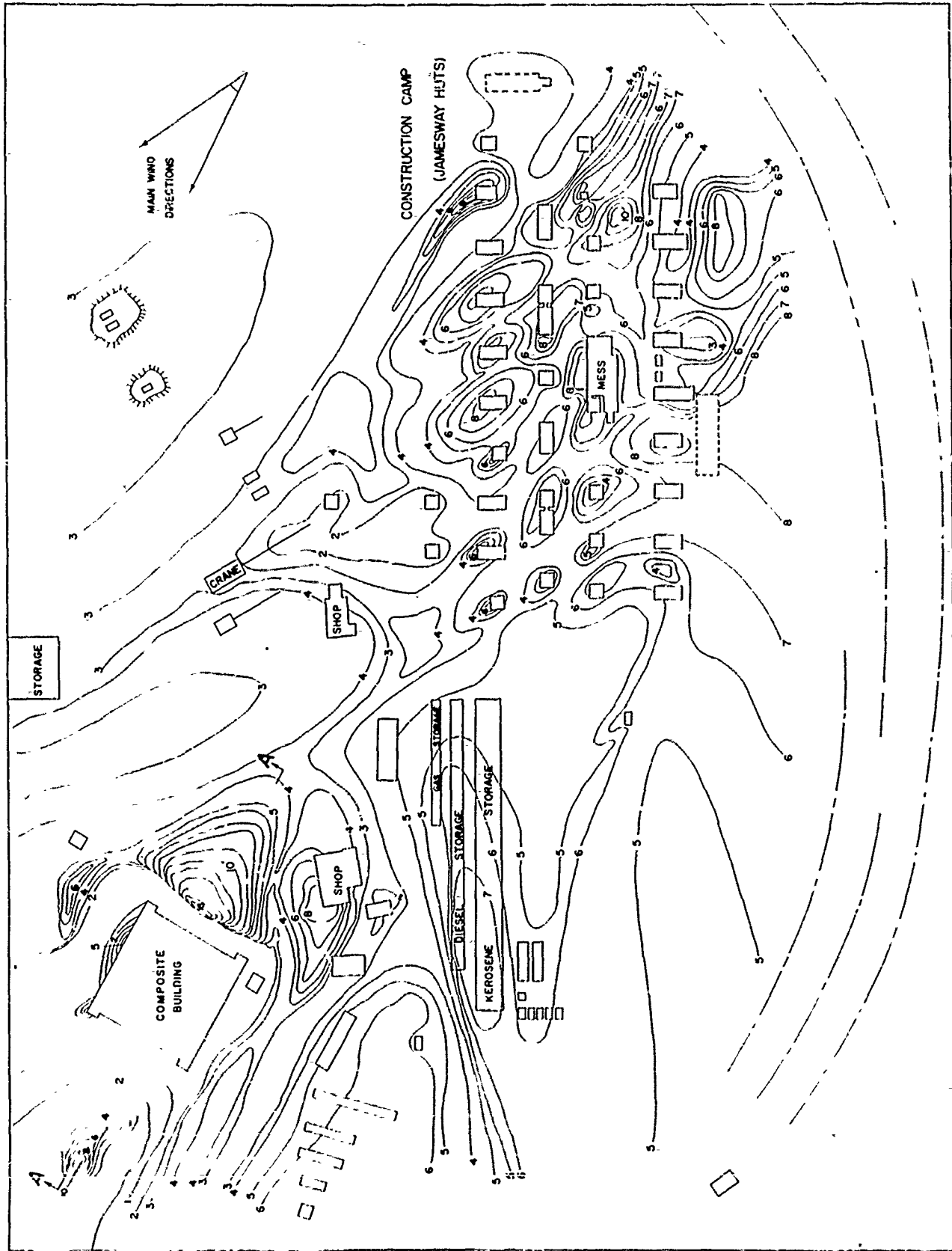


Figure 60. Dye 3, a DEWline station on the Greenland Ice Cap. This large structure is elevated 19 ft above the snow surface on extensible columns.

period, during which time snow accumulation of about 1 m/year (3 ft/yr) was expected. When the stations were designed there was a widespread belief that the most serious snowdrifts were caused by intercepting the heavy concentrations of snow in the lowest layers of the air, and the elevation of the station was expected to go far in eliminating such snowdrifting. However, the structures did generate large snowdrifts, presumably because of general air flow disturbance, with consequent influence on turbulent diffusion of snow grains. Figures 61 and 62 show, by contour plans and longitudinal profiles, the drifts formed at Dye 2 and Dye 3 during the first winter following construction. Figure 61a shows also the drifts formed by small outlying buildings and supply heaps. After the first winter bulldozers were used repeatedly to move the snowdrifts and spread the spoil over a wide area, so that it is not possible to say with certainty what might have happened if the drifts had been left to develop naturally. There are, however, some indications that they might have developed a stable profile while an open wind scour remained beneath the building.

As an alternative to building under the surface or building on stilts, it has been suggested that polar camps might be assembled from movable buildings which rest on skids; the buildings would be moved a short distance each summer to escape the drifts of the previous winter. Many problems with this procedure can be foreseen and, furthermore, experience with skid-mounted garages at the Dye sites has demonstrated the difficulty of moving buildings which are embedded in high-density, age-hardened snow.

While snowdrifting on polar snowfields presents the most difficult problems, the over-all economic consequences of drifting are greater in more habitable regions subject to seasonal snow cover only. One of the less obvious results of snowdrift formation is disturbance of the thermal regime in the ground below, which may lead to changes in water content (and hence bearing capacity) and ground water movement.



(a) Contour plan of the snow surface. Contour interval 1 ft.

Figure 61. Snowdrifts formed by the radar station Dye 2 and associated buildings during the first winter following construction. (From U. S. Army Corps of Engineers surveys and drawings.)

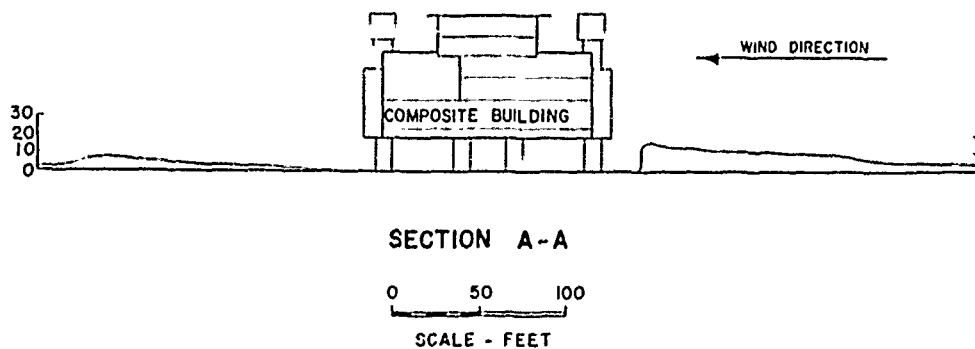


Figure 61 (Cont'd). (b) Longitudinal profile (approximately parallel to the wind direction) of drifts at the main structure, Dye 2.

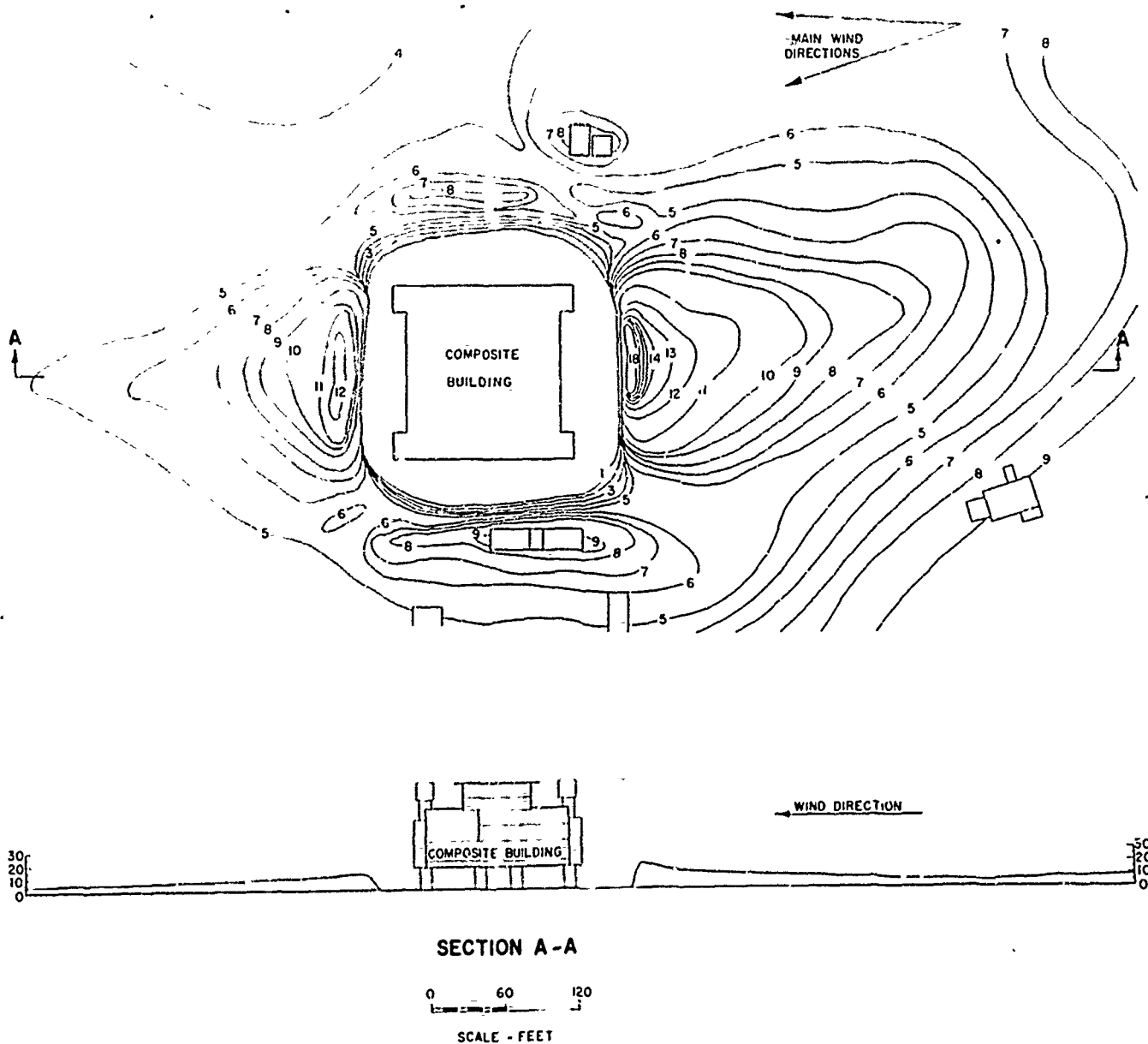


Figure 62. Snowdrifts formed by the radar station Dye 3 during the first winter following construction. (From U. S. Army Corps of Engineers surveys and drawings.) Contour plan (1-ft interval) of the snow surface and longitudinal profile (approximately parallel to the wind direction) of drifts at the main structure.

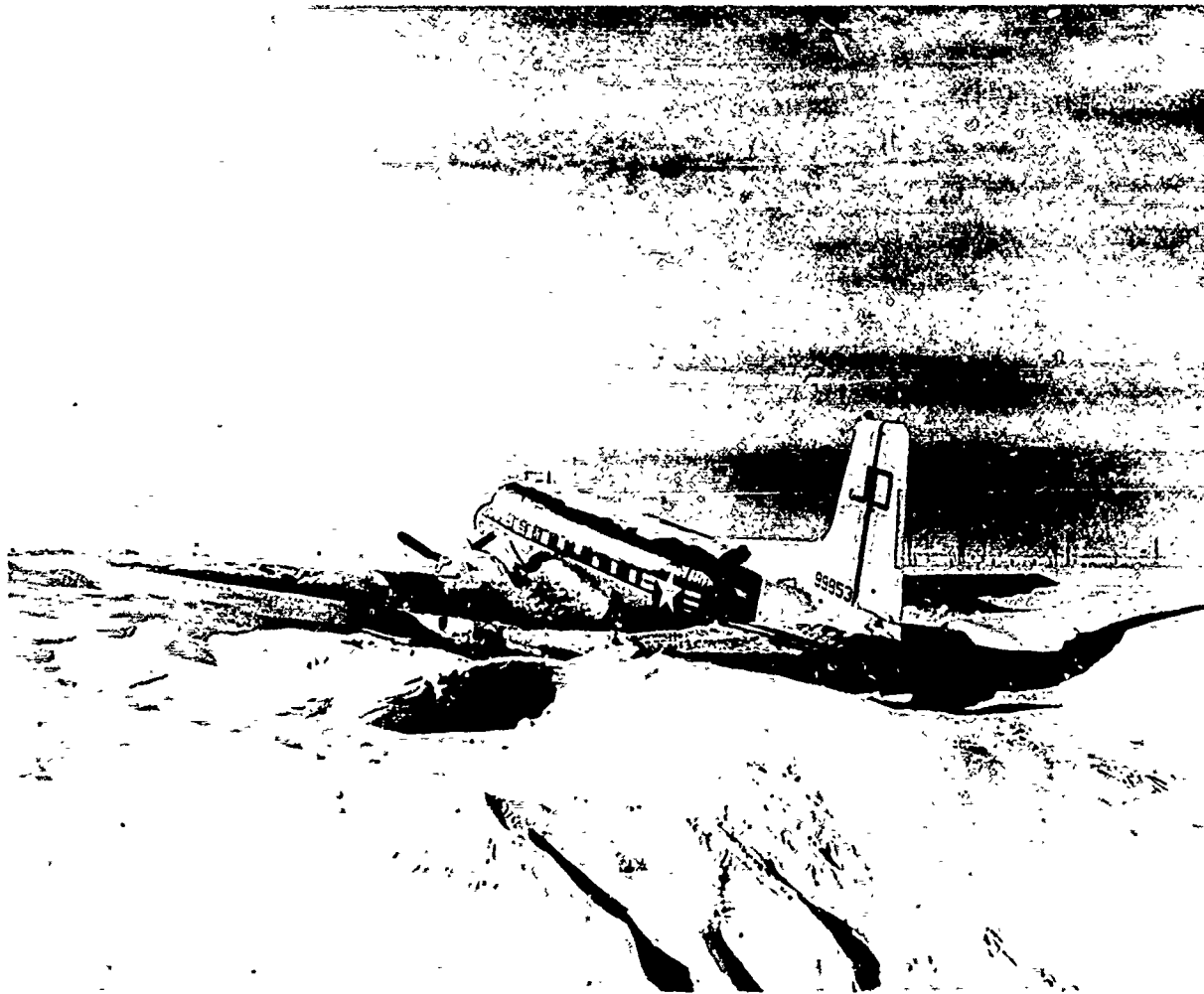


Figure 63. Snowdrifts generated during the winter by an aircraft parked in the three-point attitude with its nose into wind. Although relatively streamlined this machine formed large drifts. Much of the drift-forming turbulence can be attributed to the "stalled" attitude, but even in the level-flight attitude it seems likely that the machine would have created drift-forming eddies.

At established settlements snowdrifting is combated largely by snow fences and conventional mechanical snow removal, but where completely new installations are planned there is obviously scope for limitation of snowdrifting by suitable design and grouping of buildings. In many cases, of course, snowdrifting considerations will be subordinate to other architectural and planning requirements, but in the Russian far north, where planning for new construction is subject to a high degree of central control, it appears that drift amelioration is given serious attention. Stepanov⁸⁶ has put forward recommendations for the layout of new settlements in some detail, basing his advice on experience in the Norilsk region. The ideas put forward do not appear to have been tested yet and, since they seem to be based on a somewhat naive interpretation of the principles involved, they should be regarded with caution.

It seems clear that all but the most elaborately streamlined structure must create drifts (Figure 63 shows drifts around an airplane parked facing the wind); perhaps the most that can be done for the present is to cause drifts to occur where they can best be tolerated.

Electrical effects in blowing snow

In general, snow crystals in the air are electrically charged. When the snow is blowing, the charges on the particles may become quite large (~ 1 esu/g) and they tend to increase with increasing wind speed and decreasing temperature. The electrical field of the atmosphere is drastically modified, usually by increase of the normal positive potential gradient. Conductors exposed to the blowing snow become strongly charged, and spark discharges occur where there are large potential differences. Of particular concern is the charging of radio antennas, since the repeated discharges are a source of serious radio noise at the lower radio frequencies.

Although numerous observations have been made on charged snowflakes and blown particles, the reports in the literature are confusing at first sight (see monograph III-A1 of this series for a brief review and bibliography). Recently, however, Latham⁵¹ has made an orderly review of selected references by interpreting the reports in the light of a charge separation effect which he feels is the significant one for blizzard electrification (there are other ways in which charge separation may occur in ice). Latham's theory is based on his earlier finding, made jointly with B. J. Mason, that temperature gradients in ice, or temperature differences at ice-ice contacts, produce a separation of charge in the ice. This effect is attributed to the facts that: (i) ion concentration in ice increases rapidly with temperature, (ii) mobility of the H^+ ion is considerably greater than that of the OH^- ion. Hence the concentration gradient set up by the temperature gradient produces ion diffusion initially in favor of the protons, and the cold end of the ice develops an excess of positive charge. In a blizzard, temperature gradients in snow particles are thought to be produced by evaporation, turbulent mixing, and perhaps "asymmetric rubbing" (differential frictional heating resulting from sliding contact between ice surfaces of different size and shape). Minute splinters of ice are detached from the extremities of blown snow particles by collisions and by aerodynamic drag, and Latham^{51,52} has shown that these tiny fragments are positively charged while the larger remnant particles retain a net negative charge. It was found that when snow particles were dislodged from a snow deposit by an air stream the larger particles invariably carried positive charge, but the sign of the charge on the smallest particles varied according to the sign of the temperature difference between the air stream and the snow deposit. With air colder than the snow deposit the small particles were negatively charged, but with air warmer than the snow the small particles were positively charged.

The measurements made during real blizzards so far have been of an exploratory nature and, since systematic controls have been somewhat lacking, there are few general statements which can be made with much confidence. The principal published study on charging of antennas was made by Barré⁶, but there is new work in progress under the direction of U. Radok, Melbourne University.* The following remarks summarize the indications of available data.

In a general snowstorm, when snow is falling with strong winds, there is usually an increase in the positive potential gradient near the surface, the magnitude of the space charge at any level increasing with particle concentration and with some power of the wind speed. Exposed antennas at heights of 1 to 2 m usually become positively charged, and the charging rate varies with height⁶. Since charging rates for antennas might be expected to be small both at great height and very close to the surface, there is probably a level at which charging rate reaches a maximum. Charge magnitude increases as temperature decreases; the relatively small charge at temperatures above $-10^\circ C$ has been attributed to greater ductility (fracture resistance) and adhesion of the snow particles⁵².

When snow is picked up and blown across the surface in the absence of snowfall, antennas set at 1 to 2 m height frequently develop negative charge⁶. This is in accord

*For a chronological summary of early work, beginning with Simpson's investigations, see ref 61.

with the several observations that larger particles broken from a snow deposit are always negatively charged (see monograph III-A1). Barré's data showed that the higher of two antennas usually developed more strongly negative charge; the reason for this effect is not immediately obvious.

The chief factors in electrical stratification of blizzards are probably velocity gradients, concentration gradients, and variation of mean particle size with height. Antenna discharge occasionally indicates brief polarity reversals, perhaps resulting from extreme gusts and lulls in the wind.

Studies of radio noise during blowing snow conditions have clearly shown that there is an inverse relationship between noise level and radio frequency, noise level decreasing as frequency increases through the normal communication range. The changeover to VHF and UHF for aircraft communications has gone far toward removing the operational problem of precipitation static, but for ground stations which must still use low frequencies for long distance communication blowing snow remains a nuisance, particularly in the polar regions.

Light transmission and visibility in blowing snow

When a collimated beam of monochromatic light shines through a suspension such as falling or blowing snow, each element of the suspension, in which there may be absorption and scattering, removes from the beam an amount of flux proportional to the flux upon that element, so that

$$dF = -\sigma F dr \quad (43)$$

and hence, by integration

$$F = F_0 e^{-\sigma r} \quad (44)$$

where F is flux at distance r from an origin at which the flux is F_0 , and σ is an extinction coefficient. When visible radiation is attenuated by falling and blowing snow, absorption must be much less important than scattering, so that for practical purposes the extinction coefficient is equivalent to a scattering coefficient. The term extinction coefficient is retained here for conformity, although physical arguments are based on scattering processes.

As far as can be ascertained from a search of the literature there has been no substantial study of light transmission through falling and blowing snow, so that information has to be gleaned indirectly. For aerosols in general, the scattering coefficient for light of a given wavelength is equal to the product of particle concentration and scattering cross section of the particles (Middleton⁶³):

$$\sigma = N K a \quad (45)$$

where N is the number of particles per unit volume of air, a is the cross-sectional area of a particle, and K is a property called the scattering area ratio, i. e., area of wave front affected by a particle divided by the actual particle area a . When snow is blown by strong winds it appears that at heights above 1 m or so there is little variation of either particle size or particle shape (see p. 9), so that K and a must be fairly constant. Thus, we may expect a linear relation between σ and snow density (mass of snow per unit volume of air). When snow is falling gently the crystal size and shape may vary greatly, from small prismatic types to intricate dendrites, depending on weather conditions. In these circumstances there may be considerable variations in K and a for any given value of N , and one would not necessarily expect a simple relation between σ and snow density.

The only relevant item of information found in the literature is derived from the calibration of a blowing snow meter⁵⁰. * The experimental results show that light of

*See p. 14 for a description of this device.

about 0.6μ wavelength was attenuated by blowing snow such that

$$I_0 - I_\gamma \propto \gamma^{0.87} \quad (46)$$

where I_0 is transmitted intensity with no blowing snow and I_γ is transmitted intensity at the same distance with density of blowing snow γ .

Perhaps of more practical consequence than the attenuation of transmitted light is the reduction of contrast, and hence the limitation of visibility, by blowing snow.

A distant object is distinguished by the eye largely because of a difference in luminance, or brightness, between the object and its background, i. e., visibility depends on apparent contrast \underline{C} defined as

$$\underline{C} = \frac{B - B'}{B'} \quad (47)$$

where B and B' are the luminances of the object and its background respectively. The theory for alteration of apparent brightness when light is scattered by aerosols, first developed by Koschmieder in 1924, has been reviewed by Middleton⁶³; only simple results will be mentioned here, and the underlying assumptions will not be stated explicitly.

When the eye, or a photometric device, is directed to a perfectly black distant target, some illuminance is received due to light scattered back by aerosols. Integrating the differential equation derived from a consideration of scattering in an element of the cone of vision, and evaluating the constant from boundary conditions, it is found that, for a uniformly illuminated field,

$$B_b = B_h (1 - e^{-\sigma r}) \quad (48)$$

where B_b and B_h are the luminances from the black target and the horizon sky respectively at a distance r . Expressed another way,

$$\underline{C} = -e^{-\sigma r}. \quad (49)$$

For a target which is not perfectly black, or which is self-luminous, the expressions can be modified by adding a transmission term to eq 48. The general expression for contrast with the horizon sky then becomes

$$\underline{C} = C_0 e^{-\sigma r} \quad (50)$$

where \underline{C} is the apparent contrast at distance r , and C_0 is the inherent contrast of the target, defined by

$$C_0 = \frac{B_0 - B_h}{B_h} \quad (51)$$

where B_0 is the luminance for $r \rightarrow 0$. C_0 may take values from -1 (black body) up to large positive values.

It has been found that under daylight conditions the average human eye can distinguish a target of adequate size (say greater than 1° visual angle) if the apparent contrast exceeds 0.02. The distance at which apparent contrast reduces to this value (called the liminal contrast) is the visual range. If the target sighted is black ($C_0 = -1$) and the liminal contrast is 0.02, the limit of visibility, or "meteorological visual range" \underline{V} , can be related to the effective extinction coefficient for the line of sight σ_0 :

$$\underline{V} = 3.912/\sigma_0. \quad (52)$$

BLOWING SNOW

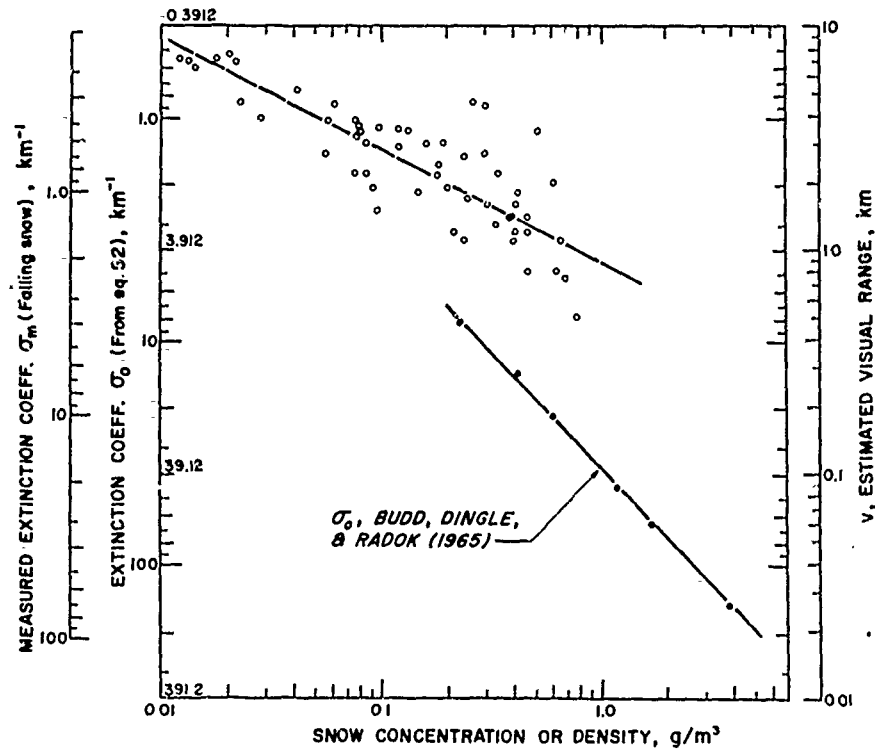


Figure 64. Visual range and extinction coefficient as functions of density for falling and blowing snow. The left-hand scale gives also an extinction coefficient σ_0 calculated on the assumption of liminal contrast 0.02 and inherent target contrast -1, and also gives a scale which corrects the data for falling snow in accordance with control measurements. (Blowing snow data from ref 12.)

The only known observations which relate blizzard visibility with blowing snow density are those reported by Budd, Dingle and Radok¹². Visual range at an Antarctic station was estimated from a line of flagged stakes and the concentration of blowing snow was simultaneously metered to give the results shown in Figure 64. Since slender stakes with small flags subtend only a small horizontal angle at the eye, liminal contrast probably varies with distance (see Middleton⁶³ for details) and the visual range as estimated may be somewhat smaller than the meteorological visual range. Nevertheless, in the absence of any other data, approximate values of σ_0 may be derived from these results. Since the dimensions of the snow particles were also determined (Fig. 7) a value of the scattering area ratio K corresponding to these σ_0 values can be calculated. For equant, sub-angular grains of "effective diameter" 0.087 mm this value of K turns out to be about 2.7 or, if 0.087 mm is the diameter of a spherical particle, 1.8. This may be compared with the limiting value of 2 found from the Mie electromagnetic theory for spherical water droplets of similar size. With bigger and blacker targets than the stakes used for the observations a smaller value of K might have been found.

Liljequist⁵³ measured visibility as a function of wind speed when snow was blowing across an Antarctic snowfield (Fig. 65). Relating wind speed to snow density according to Figures 12, 16 and 17, which refer to a location similar in essential respects to Liljequist's site, the data of Figure 65 prove to be in good agreement with those of Budd et al. (Fig. 64) for snow densities up to 5 g/m³; at higher concentrations (or wind speeds) Liljequist's data fall below the linear relation of Figure 64.

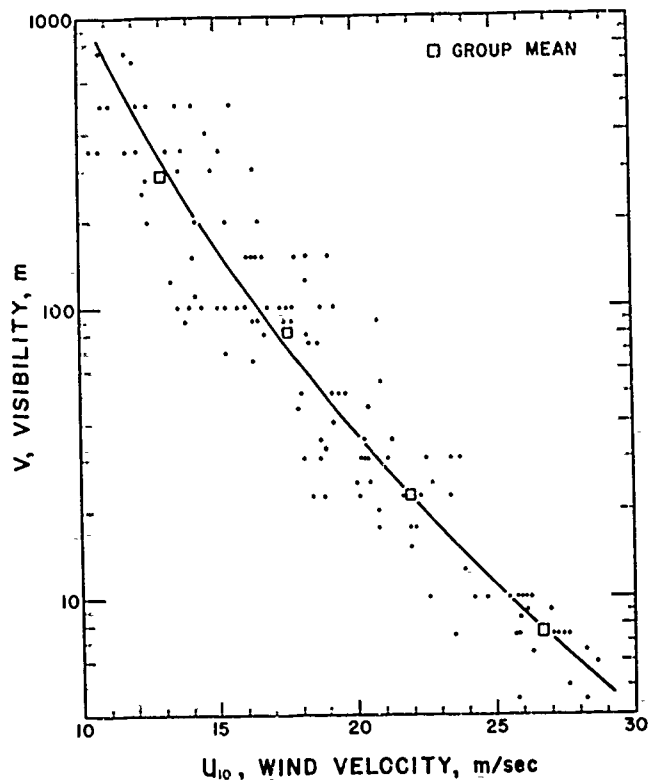


Figure 65. Logarithm of visibility as a function of wind speed at 10-m height for blowing snow conditions. (From Liljequist⁵³).

The author has measured visual range as a function of snow density on occasions when snow is falling gently (little wind), and some preliminary results are shown in Figure 64. As a control for the visual observations, contrast is measured as a function of distance with a telescopic photometric device. Since size and shape of falling crystals vary considerably from snowfall to snowfall, it is no longer permissible to regard γ as equivalent to N , as seems justifiable for the relatively uniform blowing snow. Equation 45 can be rewritten if a characteristic linear dimension of the typical snow particle, l , is defined as the cube root of the particle volume. Since $N = \gamma/l^3 \gamma_i$ and, with some reservations, $a = l^2$, where γ_i is the density of ice, then eq 45 becomes

$$\sigma = \frac{K}{l} \frac{\gamma}{\gamma_i} \quad (53)$$

Thus if K does not vary appreciably over a restricted range of optically large particle sizes, σ might be expected to be inversely proportional to particle size. Although the snowfall data have not yet been analyzed, it is noticeable that the points in Figure 64 suggest a maximum slope of about $\frac{2}{3}$ rather than unity. Rearranging eq 53, and assuming K constant, there is an implication that $(1/l)(\gamma/\gamma_i)^{\frac{1}{3}}$, or $N^{\frac{1}{3}}$, does not vary significantly over the range of conditions observed, an intriguing thought which will be pursued elsewhere.

Optical experiments on suspended snow seem worth pursuing, since there is at present very little information to satisfy either fundamental curiosity or practical needs.

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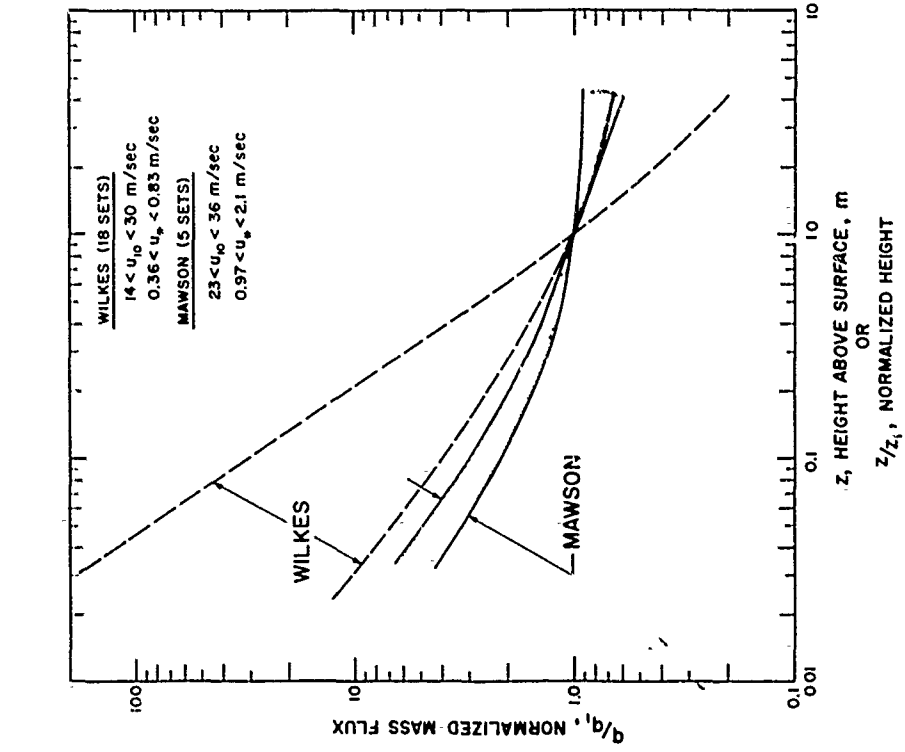


Figure A1. Limits of certain data for normalized drift density as a function of height.

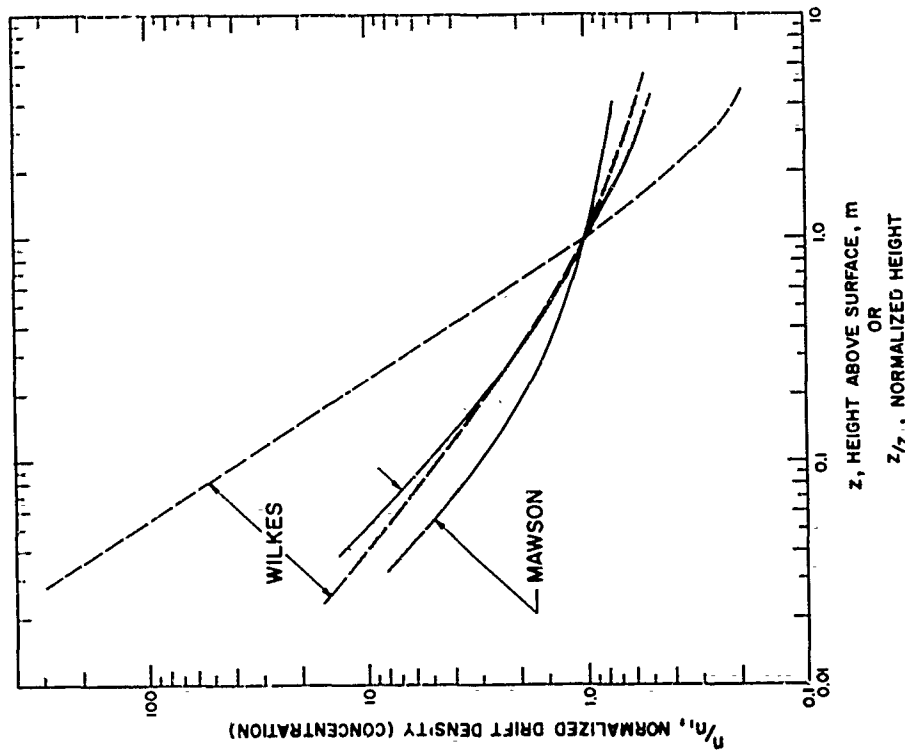


Figure A2. Limits of certain data for normalized mass flux as a function of height.

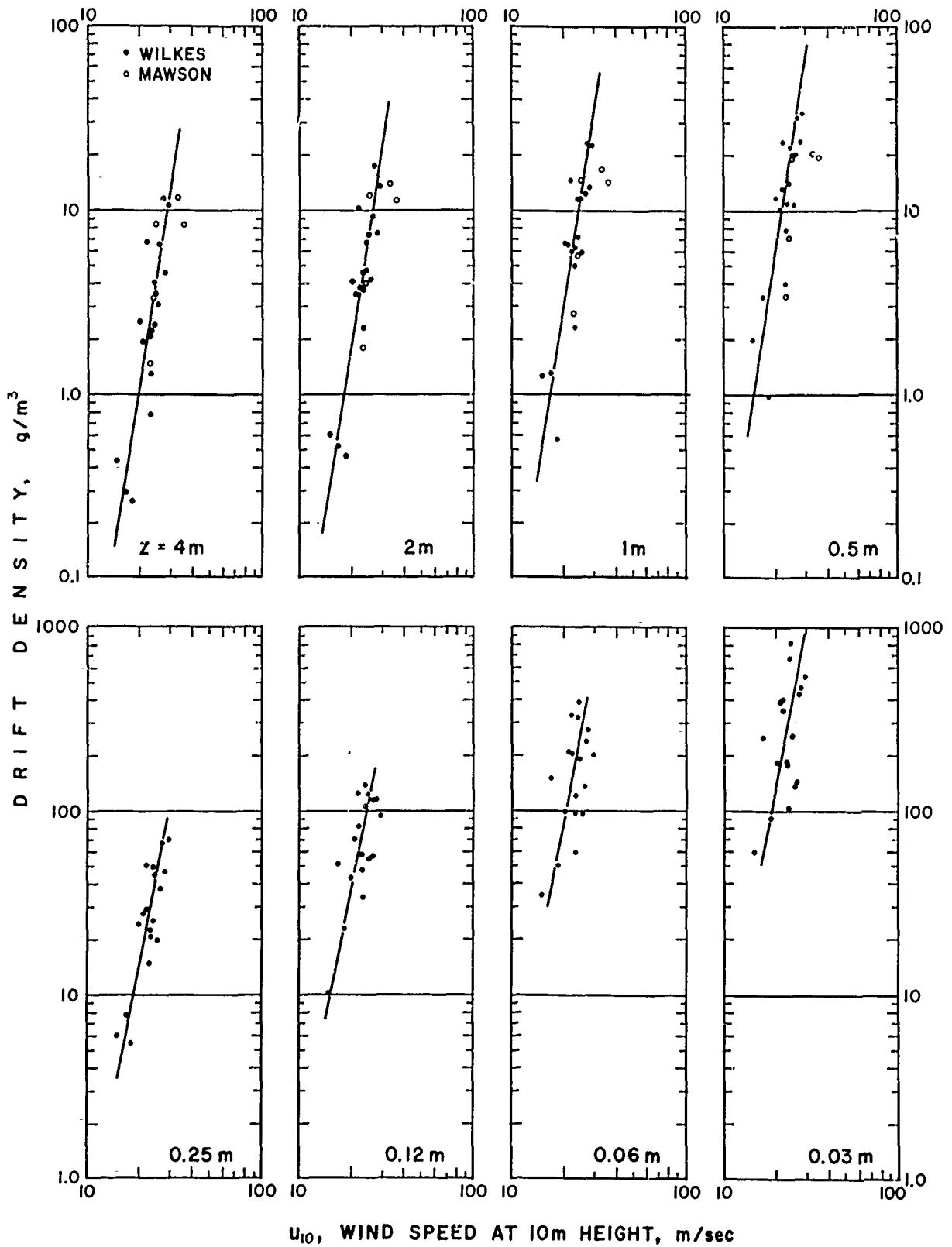


Figure A3. Drift density as a function of wind speed for a range of heights. (Wilkes data by Dingle and Radok—personal communication).

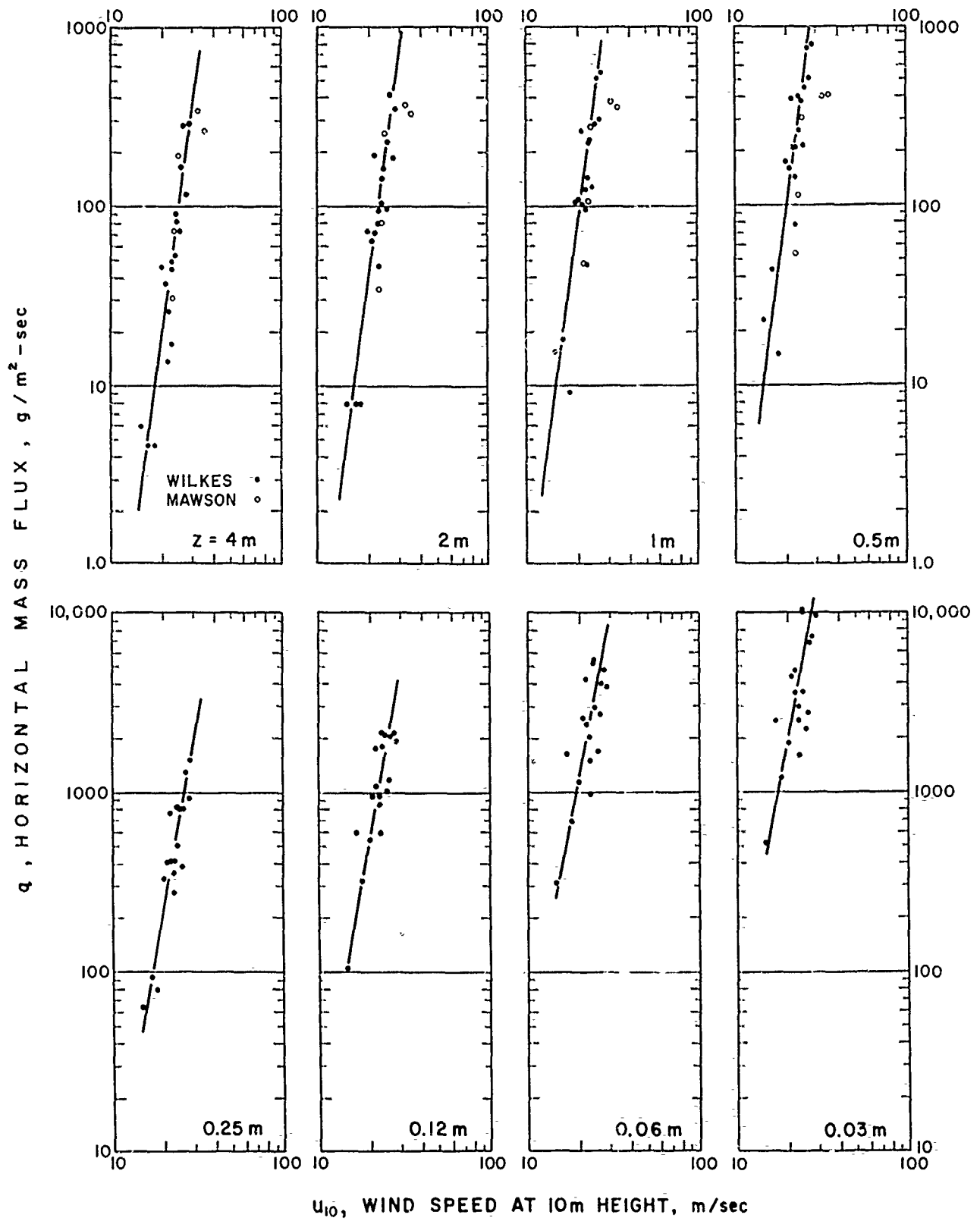


Figure A4. Mass flux as a function of wind speed for a range of heights. (Wilkes data by Dingle and Radok—personal communication).

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