

## SNOW DRIFT

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**ABSTRACT.** Wind-blown snow represents an age-old problem in the applied glaciology of most higher-latitude regions, but its physical intricacies first received attention in esoteric discussions on the long-term mass balance of polar ice sheets (Loewe, 1933, 1956). Measurements on the uniform unlimited surface of such ice sheets have shown good agreement, at least over a limited height range, with estimates for the concentration and flux of drift snow as a function of height and wind velocity based on turbulence theory. An alternative theory, developed concurrently from wind-tunnel results and field observations in Siberia, is discussed on the basis of its most recent exposition (Dyunin, 1974). Basic questions requiring further study include drift-snow concentrations at considerable heights, drift evaporation, and electrical phenomena.

The main practical aspects of snow drift relate to the prevention of excess accumulation on roads, railway lines, and avalanche slopes; and to the encouragement of accumulation in fields and forests, and other locations where frost protection and/or storage of water is desired. The methods used are reviewed; they are beginning to rely on physical concepts and theories rather than solely on empirical formulae derived from engineering experiments.

In regions of positive surface mass balance, buildings and other structures tend to become obliterated by snow drift. An important factor of this process is the fall-out of snow in the retarded flow on the windward side of the structure. Some recent attempts to measure and calculate that fall-out are discussed.

**RÉSUMÉ.** *Chasse-neige.* La neige soufflée par le vent est un vieux problème en glaciologie appliquée dans la plupart des régions de latitude élevée, mais ses implications physiques ont retenu d'abord l'attention lors de discussions esotériques sur les bilans à long terme des calottes polaires (Loewe, 1933, 1956). Des mesures sur la surface uniforme illimitée de telles calottes glaciaires ont montré une bonne cohérence, au moins dans une plage de hauteur limitée, avec les estimations du flux et de la concentration de la neige chassée en fonction de la hauteur et de la vitesse du vent selon la théorie de la turbulence. Une alternative théorique développée concurremment à partir de résultats obtenus en soufflerie et à des observations de terrain en Sibérie est discutée sur la base des plus récents travaux (Dyunin, 1974). Les questions essentielles qui demandent des études plus poussées portent sur la concentration de la neige chassée aux grandes hauteurs, l'évaporation du drift et les phénomènes électriques.

Les principaux aspects pratiques du chasse-neige concernent la prévention des accumulations excessives sur les routes, les lignes de chemin de fer et les pentes à avalanches; ainsi qu'à l'encouragement à l'accumulation sur le terrain et en forêt ou autres endroits où la protection contre le gel et/ou le stockage de l'eau sont souhaités. On passe en revue les méthodes employées, elles commencent à se rattacher à des conceptions physiques et à des théories plutôt qu'uniquement à des formules empiriques issues de l'expérience des ingénieurs.

Dans les régions où le bilan de masse en surface est positif, les bâtiments et autres structures tendent à être effacés par la chasse-neige. Un important facteur de ce processus est l'accumulation de la neige dans l'écoulement ralenti du côté sous le vent de la structure. Quelques récents essais pour mesurer et calculer ces surcharges sont discutés.

**ZUSAMMENFASSUNG.** *Schneefegen.* Treibschnee ist ein uraltes Problem in der angewandten Glaziologie der meisten Gebiete höherer Breite, doch erst in esoterischen Diskussionen um den langzeitigen Massenhaushalt polarer Eisschilde (Loewe, 1933, 1956) fanden seine physikalischen Verwicklungen die nötige Beachtung. Messungen auf der gleichförmigen und unbegrenzten Oberfläche solcher Eisschilde zeigten gute Übereinstimmung, zumindest in einem begrenzten Höhenbereich, wobei Menge und Konzentration des Treibschnees auf der Basis der Turbulenztheorie als Funktionen der Höhe und der Windgeschwindigkeit abgeschätzt wurden. Eine alternative Theorie, entwickelt aus Windkanalergebnissen und Feldbeobachtungen in Sibirien, wird auf der Grundlage ihrer neusten Darstellung (Dyunin, 1974) diskutiert. Zu den Grundfragen, die weiterer Untersuchung bedürfen, gehört die Treibschneekonzentration in beträchtlichen Höhen, die Schneeverdunstung und elektrische Phänomene.

Die wichtigsten praktischen Probleme des Schneefegens beziehen sich einerseits auf die Verhinderung übermäßiger Akkumulation auf Strassen, Eisenbahnlinien und Lawinhängen, andererseits auf die Verstärkung der Akkumulation auf Feldern, in Wäldern und anderen Stellen, wo der Schutz vor Frost und/oder die Speicherung von Wasser erwünscht ist. Die angewandten Methoden werden umrissen; sie beruhen immer mehr auf physikalischen Vorstellungen und Theorien als nur auf empirischen Ansätzen, die von ingenieurtechnischen Versuchen ausgehen.

In Gebieten mit positiver Massenbilanz an der Oberfläche sind Häuser und andere Bauwerke von der Zerstörung durch die Schneedrift bedroht. Ein wesentliches Gefahrenmoment ist dabei die Ablagerung von Schnee infolge der Verzögerung des Luftstromes an der windzugewandten Seite des Bauwerkes. Einige neuere Versuche zur Messung und Berechnung dieser Ablagerung werden diskutiert.

## 1. INTRODUCTION

The transport of snow by the wind plays a prominent part in many problems of applied glaciology. Two main problem regions may be distinguished. In *mountains*, drift-snow forms a major nourishment of glaciers and a prime ingredient of avalanches; drift-snow accumulations also have potential or real value as a form of hydrological storage. On the featureless *plains* of northern Eurasia and America, drifting snow creates problems for road and rail traffic and, by eroding a heat-insulating surface cover, for agriculture. Furthermore, surface structures tend to create snow accumulations which can interfere with their use or even bury them completely. This is an acute problem on polar ice sheets where the supply of dry loose snow is virtually unlimited and snow drift an incessant phenomenon.

The greater simplicity of the air flow over a featureless surface has made it easier to gain an understanding of basic snow-drift processes there rather than in the mountains. As a result, most systematic work on snow drift has been carried out on the Siberian plains and on the polar ice sheets of Greenland and Antarctica. However these studies have emphasized different aspects. In Siberia, snow is a seasonal problem and has practical interest. On the polar ice sheets the persistent drift-snow forms a potentially significant part of the mass balance and perhaps of their very origin. Bergeron (1965) has argued in detail how a non-linear increase in snow-drift flux with wind velocity, suggested by all observational and theoretical models (cf. Section 3 below), could account for the observed precipitation regions and the thick ice of central Greenland and Antarctica. But this has been questioned by Loewe, who first formulated the quantitative aspects of snow drift from his own observations during the Wegener expedition to Greenland (Loewe, 1933), and for Antarctica from his own observations during the second expedition of Expéditions Polaires Françaises to Terre Adélie (Loewe, 1956), before combining all the available evidence in the most authoritative estimates to date of the drift-snow losses from Greenland and Antarctica (Loewe, 1970).

These results will not be discussed here since they are somewhat outside the framework of "applied" glaciology; nevertheless they have been the driving force behind the search for a theory of drifting snow that can do more than provide rough estimates for the quantity of snow likely to be collected by a snow fence in prescribed climatic conditions. Any review of snow drift as a problem in applied glaciology must take note of the somewhat esoteric polar studies because they have provided the keys to such more special questions as visibility, electrical phenomena, evaporation, etc.

The reviewer's task is complicated by the fact that the subject has developed in two largely independent streams, reported in different languages (English and Russian). Thus even a simple comparison of results becomes a laborious undertaking, and disputes over subtleties, intractable. Nevertheless, the merging of the two streams for the benefit of applied glaciology seems an urgent necessity, and the present review makes one step in that direction.

A starting point is provided by two monographs covering all its aspects (Dyunin, 1963; Mellor, 1965). A further source of great value is a bibliography compiled by Gold (1968). These publications have made it possible to devote this review mainly to more recent advances.

## 2. DRIFT PROCESSES AND THEIR MEASUREMENTS

Two distinctly different processes are involved in the drifting of snow. In weak winds the snow particles glide or skip along the surface and only here and there are concentrated into streams deflected upward by surface irregularities. This "saltation" mode of motion has been beautifully recorded on film by Kobayashi (1972) at velocities of around 5–6 m/s at the 1 cm level (cf. Fig. 1). Quantitative analysis of such film records shows the snow concentration in the lowest millimetre above the surface to be around 1 particle per cm<sup>3</sup> and to decrease, at these wind velocities, to one hundredth of that value at a height of 1 cm above the surface.

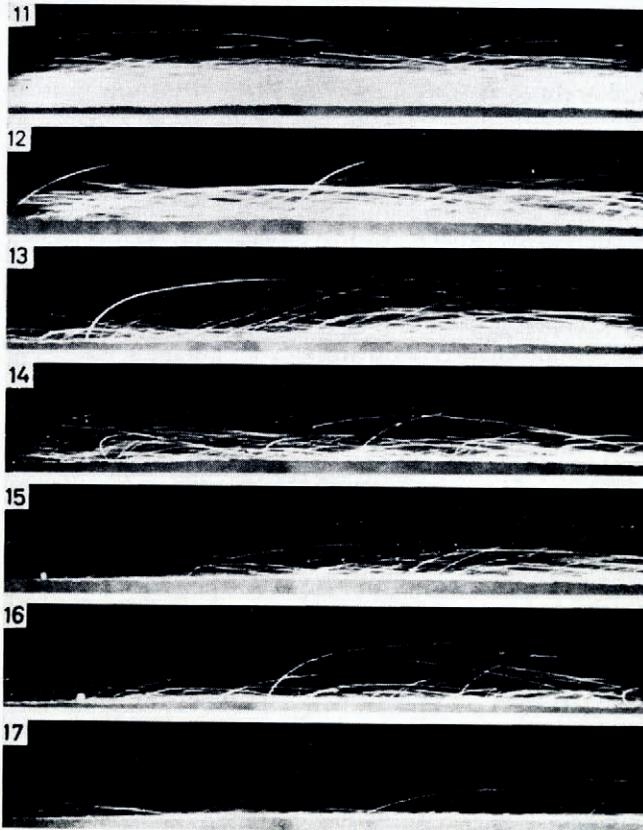


Fig. 1. Ciné pictures of snow saltation (Kobayashi, 1972).

When the wind velocity increases above a threshold which depends on the nature of the snow surface, and perhaps other factors such as temperature, the snow drift begins to go into a state of suspension by turbulent diffusion. Threshold velocities have been quoted by various authors as generally falling within the range from 8–16 m/s. At larger velocities the layer of the atmosphere with drift-snow takes on the appearance of dense fog, and snow plumes can rise hundreds of metres even over level ground. Such snow drift typically is not uniform but pervaded by snow “trombes”, eddies with vertical axes, which were described already by Wegener (1911) and are an impressive sight under special illumination. The reviewer recalls a night near the coast of Terre Adèle in 1952 when from the ship the snow drift against the bright sky above the dark outline of the ice sheet, some 15 km away, resembled protuberances during a solar eclipse. An incessant sequence of snow trombes seemed to grow to heights of a hundred or more metres in a matter of seconds and remained visible for many tens of seconds while slowly moving down the ice slope.

More commonly observed are localized drift clouds associated with topographic flow disturbances or internal flow instabilities akin to hydraulic jumps. Their practical significance outside the polar setting is to underline the large height range in which appreciable snow quantities can be transported by the wind. This clearly makes the problem of measuring the drift flux much more difficult.

For the measurement of drift-snow flux a variety of mechanical traps have been used since Andrée (1886) made the first measurements on Spitsbergen during the first Polar Year. A systematic comparison of the collection efficiency of different types of mechanical drift traps was reported by Budd and others (1966). The majority of existing drift measurements have been made with two traps: the BO<sub>2</sub> trap (cf., e.g. Dyunin, 1963, fig. 24) favoured by Soviet workers, and the rocket-shaped trap developed by Mellor (1960). A continuously recording version of the Mellor trap has been described by Tabler and Jairell (1971). An operational comparison of the efficiency of these devices in collecting polar snow in the wind range 9–14 m/s at the 1 m level showed the Mellor trap to collect 37% more, on the average, than the BO<sub>2</sub> trap. This difference must be kept in mind for any comparison of Siberian and Antarctic results.

More recently photoelectric methods have come into use for measurements of snow concentration. These do not interfere with the free air stream and provide the drift orientation directly; on the other hand, for drift flux determinations, care must be taken to combine them with suitable wind measurements (cf. Section 4).

The principle of photoelectric drift-snow gauges was discussed by Landon-Smith and Woodberry (1965) and Belov (1971); actual designs have been published by Wishart (1965) and Bird and others (1971). A particularly promising photoelectric drift gauge developed by R. A. Schmidt and Sommerfeld (1969) counts individual drift particles and measures their velocity, a quantity of basic importance for the understanding of snow-drift dynamics. According to a personal communication from Dr R. A. Schmidt, this gauge is now being used operationally in drift monitoring.

Another key factor, the particle size, has not yet been tackled with photoelectric means and presents special measurement problems in blizzards. When particles are caught directly on velvet or on slides coated with a formvar-polyvinylchloride solution the particles tend to cluster and sinter almost instantaneously, giving rise to inflated sizes. To overcome this difficulty, Dingle during the "Byrd" drift project (Budd and others, 1966) constructed a special particle trap which provided him with some 3 000 individual particles and detailed spectra for different heights. Some doubts have been raised about these in view of possible systematic differences between formvar replicas and the original particles (Dyunin, 1974), but a thorough study of replicas and their changes with time by Battye (unpublished) has revealed no such effects. Nevertheless, direct size measurements of airborne particles represent a significant need in snow-drift research, in view of the decisive role played by the particle size spectrum in the rival theories of snow drift, which will now be reviewed.

### 3. THEORIES OF SNOW DRIFT

Two main theories of drifting-snow processes have developed independently in Siberia and in Australia. The principal Soviet worker, A. K. Dyunin, has named them the "dynamical" and "diffusion" theories; their basic ideas go back to the work of Bagnold (1941) and W. Schmidt (1925), respectively. Their most telling distinction is in terms of dominant processes and vertical scales: the Siberian theory views snow drift as a near-surface phenomenon due to small eddies in the lowest 10 cm producing mainly saltation; the Australian theory (more properly described as relating to conditions on polar ice sheets) attaches the main importance to the larger eddies in the free air stream extending to tens or even hundreds of metres above the surface. The occurrence of drift particles at such heights is not in dispute, only the proportion of the total drift transport taking place in the different layers.

The only comparison to date of the two theories has recently been attempted by Dyunin (1974) almost exclusively in terms of estimates for the total transport  $Q$  as function of wind speed. Since these estimates and the underlying measurements were made with a variety of techniques, some discrepancies must be expected, and it seems essential, for a valid comparison

of the theories, to go back to their basic arguments. This is especially important for the Siberian theory, which so far lacks a critical exposition in English. Such an exposition of Dyunin's theory is attempted here on the basis of the most recent account by Dyunin (1974), which has been translated in part for the purpose.

The basic equations of the two theories are of somewhat marginal relevance from the practical point of view, and their discussion has therefore been relegated to an Appendix. Each theory results in an expression for the drift-snow concentration involving a mean eddy flux term of the form  $\overline{a''b''}$  where the variables have been assumed as sums  $a' + a''$  of a mean component  $a'$  and a deviation  $a''$ , with  $\overline{a'} = a'$  and  $\overline{a''} = 0$ . In the diffusion theory that relationship is

$$\omega n_z = -\overline{\omega'' n_z''}, \quad (\text{A.11})$$

where  $\omega$  is the particle fall velocity ( $\text{m s}^{-1}$ ) and  $n_z$  the drift-snow "density" ( $\text{g/cm}^3$ ) at level  $z$  (m). The corresponding relationship in the dynamic theory is

$$g n_z \equiv g j = \frac{\partial}{\partial z} \overline{(i'' w'' + I'' W'')}, \quad (\text{12}')$$

with  $i = \omega g n$ , and  $I, W$  being pure air-flow components for which the Appendix discussion suggests simpler alternatives. This is not a major issue since for further progress the eddy terms in (A.11) and (12') are expressed by mean flow parameters.

For this parameterization the "diffusion theory" used a gradient-flux relationship of the form

$$\overline{V_z'' n_z''} = -K_s \frac{\partial n_z'}{\partial z}, \quad (1)$$

and the working hypothesis that the eddy diffusivity for drift-snow  $K_s$  does not differ significantly from that valid in neutral conditions for momentum as well as heat and moisture, viz.

$$K_s = k u^* z, \quad (2)$$

where  $k$  is von Kármán's constant and  $u^* = (\tau/\rho)^{1/2}$  is the friction velocity.

The corresponding parameterization of the "dynamic" theory is not discussed in Dyunin's latest paper but can be found in the monograph (Dyunin, 1963). An eddy layer close to the surface is postulated where Bernoullian pressure reductions produce entrainment of snow particles. The pressure deficit has the form  $\frac{1}{2} \rho v_1^2$  and vertical gradient  $\rho v_1 (\partial v_1 / \partial z)$ . Since the horizontal velocity  $v_1$  vanishes at the surface and its gradient becomes relatively small at some quite low level, the eddy stress in Equation (A.12') and hence the drift concentration  $j$  should, on this argument, reach a maximum very close to the surface and be proportional to  $v_1^2$ . Dyunin derived a quantitative representation from a simplified boundary-layer shear profile (Dyunin, 1963, equation (128)). Two of his deduced drift-density profiles are shown in Figure 2; for comparison, a rather striking verification of the vertical density profiles predicted by the "diffusion" theory has been added from Budd's (1966[b]) study. Kobayashi's (1972) measurements at much lower wind velocities down to 1 mm above the surface do not show Dyunin's maximum which theoretically would be expected to occur at a fraction of the roughness length (Mellor and Radok, 1960).

From Dyunin's drift-density expression above it follows immediately that the drift flux close to the surface must be proportional to  $v_1^3$ . Observations and a more refined argument suggest that a threshold pressure deficit or velocity must be exceeded to break the bonds between the surface snow grains, and in due course this leads to Dyunin's well-known drift transport relationship

$$Q = 0.34(v_{0.2 \text{ m}} - 3)^3 \text{ g/m s.} \quad (3)$$

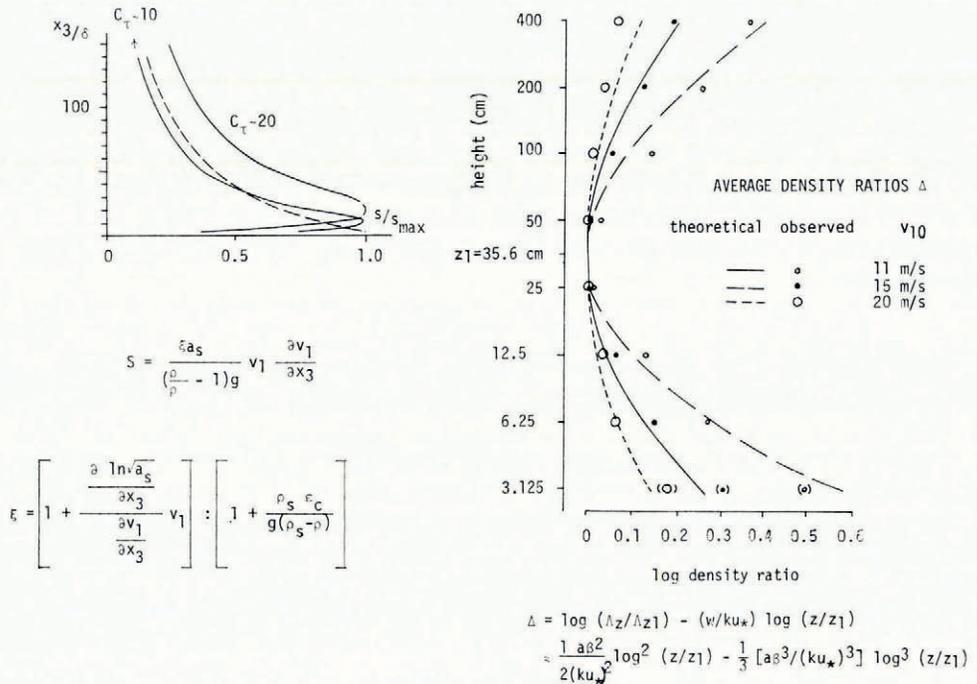


Fig. 2. Drift density profiles predicted by the "dynamic" (left) and "diffusion" (right) theories of snow drift.

By contrast, the diffusion theory of snow drift without further assumptions yields the quantitative details of the drift density as function of height or wind velocity; for constant particle fall velocity  $\omega$  these are (Dingle and Radok, 1961)

$$n_z = n_z(z/\zeta)^{-\omega/ku_*}, \tag{4}$$

and

$$n_z = n_z \exp \{[-(\omega/k^2) \log_e (z/z_0) \log_e (z/\zeta)] v_z^{-1}\}, \tag{5}$$

where  $z_0$  is the roughness length and  $\zeta$  a reference level. An interesting corollary is that measurements of  $n_z(v_z)$  over a restricted height range can be represented with practically the same precision by one of the popular power laws of the form

$$n_z = av^\alpha, \tag{6}$$

but the advantage of Equation (5) is that it predicts the values of the exponents  $\alpha$  to be expected for different levels; thus Dyunin's value of  $\alpha = 2$  would be expected, in polar conditions, near the 10 cm level, and the value  $\alpha = 5$  deduced by Liljequist (1957) from visibility data, near the 2 m level (cf. Dingle and Radok, 1961, fig. 2).

The cost of all this information is a more involved transport formula which must be found by vertical integration as

$$Q = \int_0^\infty n_z v_z dz. \tag{7}$$

Even with the assumption of a constant mean particle fall velocity  $\omega$  this gives rise to rather complex expressions; but in reality both the composition of the drift and its bulk concentration per unit mass are functions of height and wind-speed. Tractable expressions for the

change in the total drift transport with wind velocity of a single reference level have therefore frequently been constructed by fitting simple power laws to sums of partly measured and partly calculated layer transports. An example of such a formula is that derived by Budd and others (1966) for the transport through a vertical surface 1 m wide at right angles to the wind from 1 mm to 300 m above the Antarctic ice sheet at "Byrd" station,

$$\log Q = 1.18 + 0.089v_{10} \quad (Q \text{ in } g^{-1} s^{-1}). \quad (8)$$

Such simplifications have proved useful in broad assessments of the role of snow drift in the Antarctic mass balance (Loewe, 1970), and in explaining the snow accumulation up-wind of an elevated building on the Greenland ice sheet (Radok, 1968) but should not be used as evidence against the validity of the diffusion theory, as has been done by Dyunin (1974). By the same token, the diffusion theory does not invalidate the important results produced and reported by Dyunin, Kobayashi, and others from field measurements and wind-tunnel experiments on the drift flux close to the surface, in terms of net accumulation or snow collected in ditches and horizontal containers. However, the arguments here presented do show that unless one wishes to study the detailed and distinct behaviour of snow particles and of the surrounding air, the approach from the momentum equation is an unnecessary detour.

The real power of the diffusion theory comes from its detailed predictions about drift concentrations and from the verifications by direct measurements of that relatively simple quantity. For details it is necessary here to refer to Mellor (1965) and to more recent specialized accounts by Budd (1966[b]) and Radok (1970). Kobayashi's (1972) measurements confirm the theoretically predicted order of magnitude of the surface-layer drift concentration, and measurements with the photoelectric particle counter have suggested comparable drift-particle and wind velocities (personal communication from R. A. Schmidt). This further supports the diffusion theory, and makes it the essential stepping stone to the understanding of various unsolved basic problems of snow drift. Some of these will be discussed in the next section.

#### 4. SOME BASIC PROBLEMS OF SNOW DRIFT

The first unsolved problem to be mentioned concerns the vertical scale of snow transport. Transports found by extrapolation to layers above measurement levels are as large or larger than those found near the surface. Observational confirmations of the computed drift concentrations at high levels are needed to round off the otherwise satisfying verification of the diffusion theory. Such high-level measurements seem well within the capabilities of photoelectric devices. In using this type of system for estimates of drift flux during a time  $T$  at a given level,

$$\int_0^T (v_z' + z_z'')(n_z' + n_z'') dt,$$

it will be necessary to multiply the outputs of anemometers and drift-density gauges with similar time resolutions, rather than use average winds and densities whose product neglects the potentially significant eddy contribution

$$\int_0^T v_z'' n_z'' dt.$$

As a second basic problem, electric phenomena associated with snow drift have been noted since the beginning of polar exploration (e.g. Simpson, 1919-23, Vol. 1, 1921, p. 309; Sheppard, 1937) but present rather difficult measuring problems. The most advanced study to date was made by Wishart (1970) who measured bulk as well as individual particle charges

and currents created in wires strung across the wind. The wire currents were found to be substantially larger than the charge carried by the drift-snow, suggesting the additional presence of negative ions as well as space charges. Further measurements are needed to confirm these and to establish the charging mechanisms responsible for blizzard electricity; an elaborate programme has been drawn up by Wishart (unpublished) for this work. No backer has as yet been found for this but the links between blizzard and cloud electricity, the role of blizzard discharges in the terrestrial ozone balance, and the practical implications of blizzard statics for radio communications, will all in due course demand greater attention for this field of study.

The final basic problem to be mentioned may also play a role in blizzard electricity but has great practical significance in its own right. This is the evaporation of drifting snow, a well-researched though as yet largely untranslated topic in the Russian literature, which has recently also caught the attention of American workers. To do it full justice would therefore require the comparative approach of Section 3, but this will have to be done in a separate paper. Here it must suffice to mention the most important recent results and their practical implications.

The majority of these stem from an ambitious attempt by Schmidt (1972) to model all the major processes contributing to the evaporation (termed "sublimation" by him) of single drift particles. The study finds that at drift densities occurring in natural blizzards interaction effects are negligible, except possibly in the lowest few millimetres above the snow surface. The evaporation rate appears to double for each 10 deg temperature rise in the range from  $-20^{\circ}\text{C}$  to  $0^{\circ}\text{C}$  and more than double when the particle diameter is doubled. The percentage mass loss in unit time by evaporation, however, increases markedly with decreasing particle size. The evaporation rate is proportional to the saturation deficit when conduction and convection are considered alone, and doubled by allowing for solar radiation effects. Altitude and particle-shape effects appear to be of lesser importance, but further work by Lang (unpublished) has brought out the important role played by atmospheric turbulence (personal communication from R. A. Schmidt).

The theory will no doubt be extended in due course by more sophisticated treatments of the radiation transfer in snow drift and of diabatic stability conditions. Even without this, the inferred humidity and temperature profiles corresponding to steady-state steadily evaporating snow drift are of considerable importance for practical snow-control considerations (cf. Section 5 below). They also point the way towards an improved model of the katabatic wind systems of polar ice sheets which may be regulated by the interplay of adiabatic heating with radiational and evaporative cooling in the associated snow drift (Radok, 1973).

##### 5. CONTROL OF SNOW-DRIFT ACCUMULATION AND EROSION

Techniques and experiments in controlling snow-drift accumulation and erosion long preceded the more fundamental studies discussed so far and are well summarized in Mellor's (1965) monograph. Dyunin (1963) provides a great deal of additional field and wind-tunnel information on the quantitative effects of snow fences and vegetation barriers on the nearby wind field; this needs to be translated and compared with the independent, more recent experimental results of Martinelli (1973[a]) and Tabler and Veal (1971).

Martinelli (1973[b]) has summed up the features that have been found useful in practical tests. Tabler (1974) has reported in more detail on the quantitative aspects of fence design and efficiency.

According to Martinelli (1973[b]) the trapping efficiency of a fence appears to increase with fence height  $H$  up to about  $H = 3.4$  m, close to the practical upper limit. The fence "density" (solid percentage cross-section area of the fence) should be between 40 and 50%; slightly higher values have been recommended by Dyunin (1963) who uses a "transparency"

factor defined as  $(100 - \text{density})$ . The main requirement is that the fence must remain permeable; this condition also suggests the size of the bottom gap, in the range  $0.1H$  to  $0.15H$ . For a fence with bottom gap equal to  $0.1H$  experience shows that the final lee drift will vary in length from  $10H$  (100% density) to  $15-17H$  (50% density). Height and density have similarly systematic effects on the maximum depth and its location along the drift. Fences should be at least  $2H$  long and continuous; inclined and tandem fences are not significantly better.

In special contexts the relocation of a fence has been used to create drift accumulations of up to 10 m (cf. Mellor, 1965, p. 52). In principle it ought to be feasible, if perhaps not cost-effective, to design a fence which rises automatically as the snow settles behind it.

With reasonably general agreement on what makes an efficient trap for surface drift, the main current research is going into the problem of evaporative losses and their control. This topic also occupied a good deal of Dyunin's (1963) monograph, in addition to his treatment in an earlier work (Dyunin, 1961), but both await being adequately translated. In the English literature R. A. Schmidt's (1972) seminal modelling discussion was preceded by the more pragmatic approach of Tabler (1971) which has not been accessible to this reviewer but appears from later summaries, closely to parallel that of Dyunin (1963).

The principal new concept is the "transport distance"  $R_m$  (Tabler, 1971), called by Dyunin "effective length of snow-collecting basis,  $L_e$ ". This is defined as the distance over which the average drift particle is completely evaporated and differs in general from the "contributing distance" ( $R_c$  in the English literature,  $L_s$  in Dyunin (1963)). The amount of drift-snow collected by a fence is the fraction  $\theta$  of the precipitation  $P$  which is blown off this surface over the distance  $R_c$ , reduced by the evaporation loss.  $\theta$  is called "relocation ratio" by Tabler and "blow-off coefficient" by Dyunin. In these terms the snow collected becomes (Tabler, 1975[a])

$$Q_c = \theta PR_c(1 - R_c/\zeta R_m), \quad (9)$$

or (Dyunin, 1963, equation (273)).

$$Q_c = \theta PL_s - L_s I_s \Delta e, \quad (10)$$

where  $I_s$  is the evaporation rate per unit saturation deficit  $\Delta e$ . The second term on the right-hand side equals the total evaporation loss,  $\theta PR_c^2/2R_m$ .

Typical values of  $\theta$  are quoted by Tabler (1975[a]) in the range of 0.5–0.9 for Wyoming, and by Dyunin (1963, table 39, p. 274) from 0.37 to 0.88 for Siberia. The magnitude of  $R_m$  is of the order of a few kilometres and similar to that of  $R_c$  in natural condition. In order to reduce evaporation loss in areas where drift accumulation is desired, the natural contributing distance  $R_c$  can be subdivided by additional fences, but this evidently must be judged in terms of total cost effectiveness.

The foregoing approach has recently been considerably refined by Tabler (1975[a]) with Schmidt's (1972) drift evaporation model. The revised procedure is based on the particle-size distribution measured in Antarctica by Dingle and reduced to a handy mathematical form by Budd (1966[b]). Figure 3 shows schematically the amount of detail that can now be

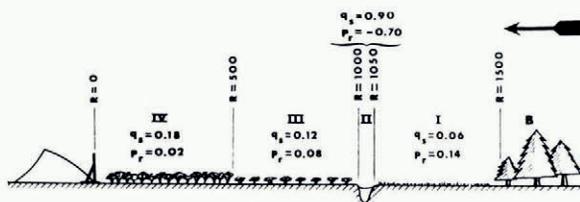


Fig. 3. The control of drift accumulation (Tabler, 1975); for details see text.

handled;  $q_s$  and  $p_r$  are the water-equivalents of snow stored and precipitated, respectively, and need to be measured or estimated. Comparison of calculated and observed drift accumulations show very satisfactory agreement, for all practical purposes.

Progress has also been made with the corresponding problems of drift accumulation by natural topographical features and by substantial structures. A very general attack, by Berg and Caine (undated), on the first problem has not yet been completely successful in predicting the shape of natural drift accumulations, but points the way to a complete solution. For practical purposes a multiple regression formula involving terrain slopes appears to have achieved considerable success (Tabler, 1975[b]). This empirical approach could be extended by means of sequences of aerial photographs taken during a melt season. Such photographs would show the most persistent drift accumulation which could then be systematically analysed in terms of mapped topographical parameters and weather records for the preceding accumulation season.

The problem of drift-snow accumulations on buildings and structures also has not yet found its definite solution. Following the progressive obliteration of several I.G.Y. stations in Antarctica, two different remedies were explored, viz. underground and elevated construction. The latter approach spawned a lot of wind-tunnel work and the exploration of the similarity laws of snow drift (Odar, 1965). A review of problems and possibilities in this area has recently been made by Norem (1975). However, it is not practicable to model basic snow processes such as sintering and cornice formation, and difficult in most wind tunnels to produce realistic boundary-layer depths of several times the height of the model building. Hence more recent wind-tunnel studies have mostly limited themselves to deducing the likely regions of snow deposition from the structure of the wind field and the surface stress (Melbourne and Styles, 1967).

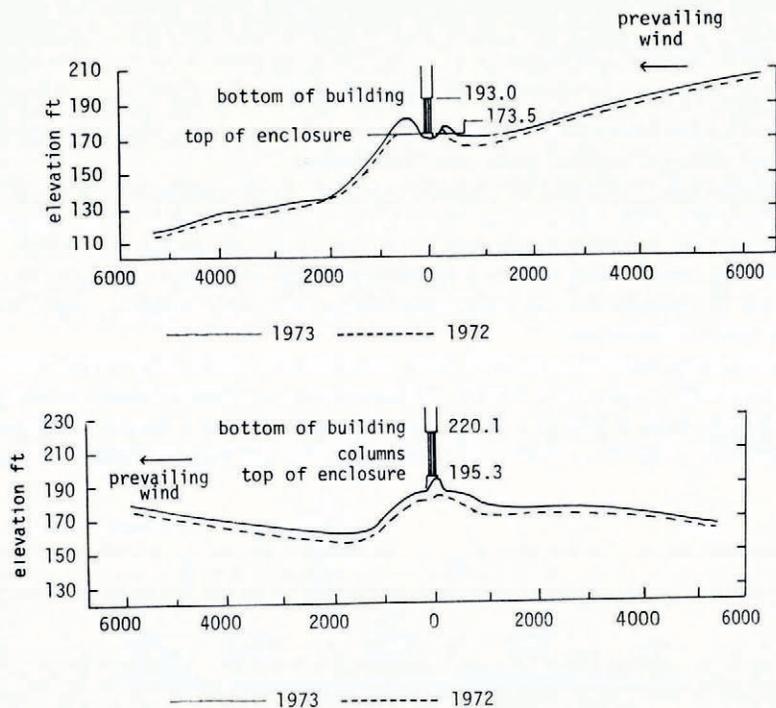


Fig. 4. The accumulation of drift snow on two radar stations in Greenland (Tobiasson and others, 1975).

Greater realism can be achieved in this context by gradually building up the tunnel surface in regions of flow stagnation (Cermak, 1966). A useful alternative has been provided by observations on relatively large model buildings set up near, or exposed to similar field conditions as, the actual buildings (Tobiasson and Reed, 1966).

A major result of these studies was to throw doubt on the original reasoning behind the elevated-construction approach. In line with the central tenet of the "dynamical" drift theory, this was that the bulk of the drifting snow remains close to the surface and should be able therefore to pass through unhindered below a raised building. Against this, Radok (1968) showed that the drift accumulation up-wind of a radar building on the Greenland ice sheet incorporated broadly all the snow-drift flux divergence in the layer between the surface and the top of the building, almost 100 ft (30 m) above the surface.

The drift accumulations on this and another radar building on the Greenland ice sheet have remained under constant surveillance (Tobiasson and others, 1972, 1975) and developed characteristic shapes at the two sites (Fig. 4). These may be due to the different surface slopes of the ice sheet; although small, such differences in slope and also in surface curvature can have very marked effects on the snow accumulation rate (Budd, 1966[a]) which might be further accentuated by an obstacle. Another example is provided by the elevated building at "Casey" station, Antarctica, where the drift accumulation each year tends to take quite different shapes along two wind flow lines only 400 ft (120 m) apart (Fig. 5). It would be interesting to apply the statistical relationship developed by Tabler (1975[b]) to these topographical situations.

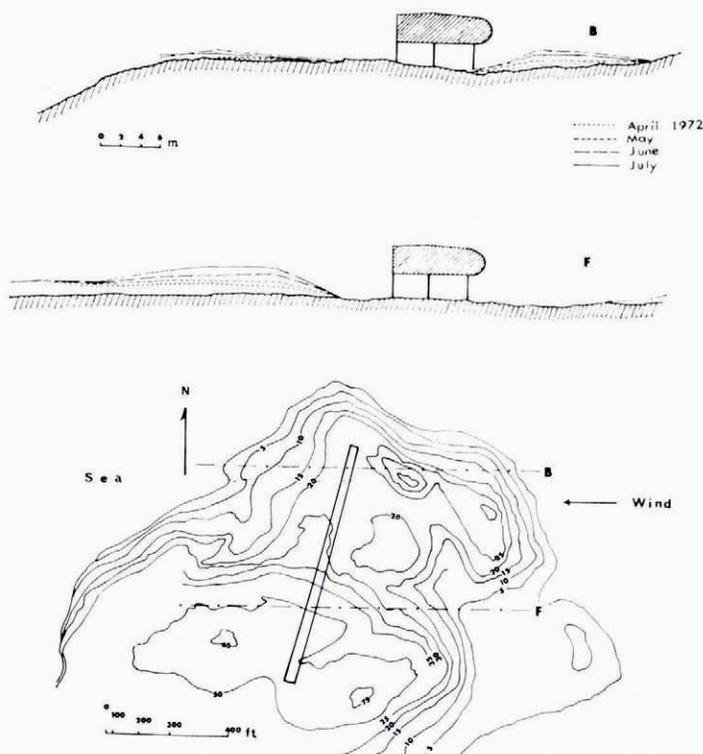


Fig. 5. The accumulation of drift snow along two wind flow lines past the station building at Casey, Antarctica (Melbourne and Styles, 1967, and Antarctic Division data).

The main practical problem in the case of elevated structures is whether especially the up-wind drift accumulation can be allowed to build up to an equilibrium shape without blocking the space below the building. An experiment is currently being carried out at "Casey" station by P. Keage of the Australian National Antarctic Research Expeditions to measure the rates of build-up along lines B and F of Figure 5 during individual blizzards. His preliminary measurements have already shown that a substantial slowing-down of the flow occurs during the approach to the building even in the layer seemingly free to move through the gap. This is accounted for by the general pressure pattern up-wind of such structures, as established by Melbourne and Joubert (1971) and schematically shown in Figure 6.

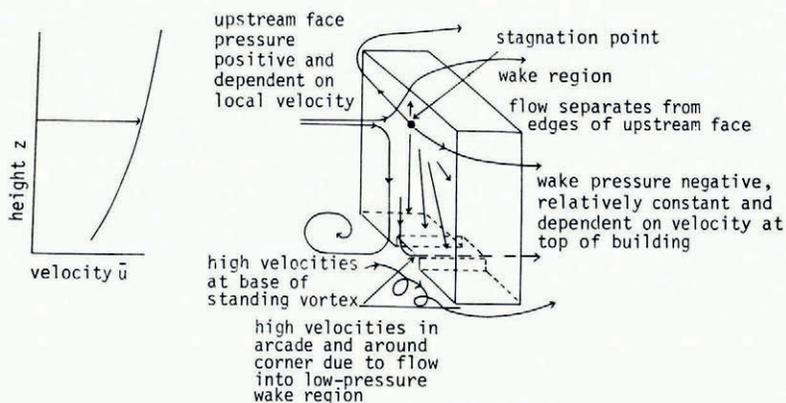


Fig. 6. The flow around a tall building with bottom gap (Melbourne and Joubert, 1971).

## 6. CONCLUSION

It has not been possible here to do more than touch on the main current activities in the field of snow-drift research and operations, and in particular to mention the diversified related work going on in mountainous regions. But the advances made with quantifying the effects of natural flow around surface obstacles, and with the understanding of drift evaporation, provide an assurance that, as a problem in applied glaciology, drifting snow is well under control in the plains. The mountains, however, are a different matter.

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## APPENDIX

## BASIC EQUATIONS OF TWO THEORIES OF SNOW DRIFT

Dyunin (1974, p. 10–13), to explain the “dynamic theory”, “proceeds from the dynamical equation for a two-phase solid flow. The vertical component of that equation for the particular case of a plane current moving along a horizontal surface has the following form

$$\frac{1}{g} \left| \frac{\partial}{\partial t} i + \frac{\partial}{\partial x} qw + \frac{\partial}{\partial z} iw \right| = s(\gamma_s - \gamma) + \frac{\gamma_s}{g} s \epsilon - \frac{\partial p}{\partial z} \frac{\partial \tau}{\partial x}, \quad (7')$$

where  $g$  is the acceleration of gravity,  $m/s^2$ ,  $i$ ,  $q$  are the mass fluxes of drift-snow in the direction of the co-ordinate axes  $z$  (vertical) and  $x$  (horizontal), respectively, in  $kg/m^2 s$ ,  $t$  is the time,  $s$ ,  $w$  is the average vertical velocity of the snow particles,  $m/s$ ,  $s$  is the volume concentration of the solid particles,  $\gamma_s$ ,  $\gamma$  are the specific weights of snow and air,  $kg/m^3$ ,  $\epsilon$  is the total acceleration of other mass forces, e.g. the force of electrical attraction and repulsion between charged particles  $m/s^2$ ,  $p$  is the pressure on a horizontal surface  $kg/m^2$ ,  $\tau$  is the shear stress along a vertical surface at the given point of the flow,  $kg/m^2$  [translated from Dyunin, 1974].

Although subsequently simplified out of recognition this equation needs a closer analysis as basis and starting point of the “dynamic theory”.

To arrive at (7') it appears that the author has started from the vertical equation of two-dimensional ( $x$ ,  $v$  horizontal;  $z$ ,  $w$  vertical) motion of a “two-phase” snow-air mixture of density  $\rho = s\rho_s + (1-s)\rho_a$ , viz.

$$\frac{\partial w}{\partial t} + v \frac{\partial w}{\partial x} + w \frac{\partial w}{\partial z} + g - \epsilon + \frac{1}{\rho} \frac{\partial p}{\partial z} + \frac{1}{\rho} \frac{\partial \tau}{\partial x} = 0, \quad (A.1)$$

where  $p$  is the atmospheric pressure,  $\tau$  the vertical shear stress,  $\epsilon$  the sum of other body forces and  $g$  the acceleration of gravity. Multiplication by the total density gives

$$\frac{\partial(\rho w)}{\partial t} + \frac{\partial(\rho w v)}{\partial x} + \frac{\partial(\rho w w)}{\partial z} + g\rho - \epsilon\rho + \frac{\partial p}{\partial z} + \frac{\partial \tau}{\partial x} = 0, \quad (A.2)$$

where the convective terms have been transformed according to

$$v \frac{\partial(\rho w)}{\partial x} + w \frac{\partial(\rho v)}{\partial z} = \frac{\partial(\rho w v)}{\partial x} + \frac{\partial(\rho w w)}{\partial z} - w \left[ \frac{\partial(\rho v)}{\partial x} + \frac{\partial(\rho w)}{\partial z} \right], \tag{A.3}$$

the last term on the right in Equation (A.3) vanishing by virtue of mass conservation when evaporation of the drift particles is disregarded.

The two "phases" of the flow can now be considered additively. Changing with Dyunin to weights instead of densities by writing  $\rho_s g = \gamma_s$ ;  $\rho_a g = \gamma$ , and distinguishing between the vertical velocities of the air ( $w$ ) and snow ( $\omega$ ), Equation (A.3), becomes

$$\begin{aligned} & \frac{\partial(s\omega\gamma_s)}{\partial t} + \frac{\partial(sv\gamma_s\omega)}{\partial x} + \frac{\partial(sw\gamma_s\omega)}{\partial z} + s\gamma_s - s\gamma_s\epsilon + sg \frac{\partial p}{\partial z} + sg \frac{\partial \tau}{\partial x} + \\ & + \frac{\partial((1-s)w\gamma)}{\partial t} + \frac{\partial((1-s)v\gamma w)}{\partial x} + \frac{\partial((1-s)w\gamma w)}{\partial z} + (1-s)\gamma g + (1-s)g \frac{\partial p}{\partial z} + (1-s)g \frac{\partial \tau}{\partial x} = 0. \end{aligned} \tag{A.4}$$

The two lines show the balances for the two phases, but only the first is really relevant to the discussion. Introducing finally Dyunin's notation

$$i = s\omega\gamma_s, \tag{A.5}$$

$$q = sv\gamma_s, \tag{A.6}$$

and dividing by  $g$  we obtain finally

$$\begin{aligned} & \frac{1}{g} \left( \frac{\partial i}{\partial t} + \frac{\partial}{\partial x} qw + \frac{\partial}{\partial z} iw \right) + s(\gamma_s - \gamma) - s\gamma_s\epsilon + s \frac{\partial p}{\partial z} + s \frac{\partial \tau}{\partial x} + \\ & + \frac{(1-s)}{g} \left( \frac{\partial w\gamma}{\partial t} + \frac{\partial(v\gamma w)}{\partial x} + \frac{\partial(w\gamma w)}{\partial z} \right) + (1-s) \left( \frac{\gamma}{s-1} + \frac{\partial p}{\partial z} + \frac{\partial \tau}{\partial x} \right) = 0. \end{aligned} \tag{A.7}$$

The first line resembles Dyunin's basic equation (7') except that there the factor  $s$  has been omitted in the pressure and stress gradient terms; moreover the introduction in the first line of an Archimedean force requires a correspondingly modified air density in the second line. Subsequent simplifications of the basic equation obliterate these differences.

Before simplifying Equation (A.7) the corresponding development in the "diffusion theory" should be noted. This can be written down directly as the continuity equation for the drift-snow mass per unit volume at level  $z$ , called "drift density" for short and denoted by  $n_z$ :

$$\frac{\partial n_z}{\partial t} + \frac{\partial(n_z v)}{\partial x} + \frac{\partial(n_z \omega)}{\partial z} = 0. \tag{A.8}$$

For the only concrete case considered in practice, that of steady-state horizontally homogeneous drift, Equation (A.8) reduces to

$$\frac{\partial(n_z \omega)}{\partial z} = 0, \tag{A.9}$$

whereas the first line of Equation (A.7) becomes approximately

$$\frac{1}{g} \frac{\partial}{\partial z} (iw) \approx s\gamma_s + s \frac{\partial p}{\partial z}, \tag{A.10}$$

if the small Archimedean and other mass forces are neglected.

As next step in both developments, the vertical drift fluxes are  $i = \bar{i}' + i''$ ,  $w = w' + w''$ , etc., and  $\bar{i} = \bar{i}'$ ,  $\bar{w} = w'$ , etc., where the bar denotes a time-space average. Applying the averaging process to Equation (A.9) then gives

$$\bar{n}_z' \omega' = -\bar{n}_z'' \omega''. \tag{A.11}$$

Dyunin's manipulation is more involved and follows again in translation; the present discussion has however established that the terms  $\partial p/\partial z$  and  $\partial \tau/\partial x$  should everywhere have  $s$  as factor:

"Equation (7') is the equation of turbulent two-phase flow averaged over space and time. Let  $q, i, w$  be mean values,  $q', i', w'$  their instantaneous values,  $q'', i'', w''$  their fluctuations

$$\overline{i'w'} = iw + \overline{i''w''}; \quad \overline{q'w'} = qw + \overline{q''w''}, \tag{8'}$$

where the bar indicates time-space averaging. The second terms on the right (the correlations between  $q, i, w$ ) do not vanish; the fluctuation  $i'', w''$  and  $i'', q''$  can have the same or opposite signs.

It follows from this that

$$p = p_M + \frac{1}{g} \overline{(i''w'')}; \quad \tau = \tau_M + \frac{1}{g} \overline{(q''w'')}, \tag{9'}$$

as is easily derived from (7'), where  $p_M, \tau_M$  are the mean stresses and the correlations represent additional turbulent stresses.

Stricter reasoning gives

$$\left. \begin{aligned} p &= p_M + \frac{1}{g} \{ \overline{(I''W'')} + \overline{(i''w'')} \}, \\ \tau &= \tau_M + \frac{1}{g} \{ \overline{(G''W'')} + \overline{(q''w'')} \}, \end{aligned} \right\} \tag{10'}$$

where  $I''$ ,  $G''$  are the fluctuations of the vertical and horizontal air flows and  $W''$  is the fluctuation of the transverse component of the air flow.

Clearly the stresses  $p$ ,  $\tau$  have complex structures due to the correlation components.

Let us consider the case of established snow drift realizing its full transport capability and maximum transport rate.

For such a drift the vertical mass flux is zero on the average. Otherwise the snow carried by the current would increase or decrease, contradicting the assumption of steadiness. With  $w$  zero on average irrespective of  $i$  the right-hand side of Equation (7') becomes zero.

Let us introduce the quantity

$$j = s(\gamma_s - \gamma),$$

which is the snow concentration in a unit mass of the snow-air mixture (allowing for Archimedean forces). In that case we find from Equation (7')

$$j = \frac{\frac{\partial p}{\partial z} + \frac{\partial \tau}{\partial x}}{1 + \left(1 + \frac{\gamma_s}{j}\right) \frac{\epsilon}{g}}. \quad (11')$$

It is natural to expect in a steady drift flow that

$$\frac{\partial \tau}{\partial x} \approx 0;$$

the values  $\gamma_s/j \approx \gamma/\gamma_s \approx 10^{-3}$ , and the other force can be neglected in comparison with gravity, as shown in (8, §8). The normal pressure  $p_M$  in Nature equals the barometric pressure and is equal over the entire thickness of the current. Therefore

$$j = \frac{1}{g} \frac{\partial}{\partial z} \{ \overline{(i''w'')} + \overline{(I''W'')} \} \quad (12')$$

Measurements of normal pressures in wind do not reveal appreciable deviations from zero in the gradient  $\partial p/\partial x$  with increasing distances from the Earth's surface. Therefore deviation from zero in Equation (12') would be expected in the immediate neighbourhood of the surfaces. This leads to the conclusion of the "surface nearness" of low drift. For very low hydraulic values  $\omega$ , particles can be lifted by the wind through the diffusivity

$$\frac{\partial \overline{(i''w'')}}{\partial z} > 0,$$

to great heights; at first glance this confirms the correctness of the diffusion theory but only for very small snow particles, which exist only for a short time, as we shall see below" [translated from Dyunin, 1974].

Dyunin's equation (12') contains a term relating to the air itself: this arises from his inclusion of the total pressure and shear-stress gradients in Equation (7'), instead of only a proportion relating to the drift phase as in the first line of Equation (A.7). The difference is however lost in the subsequent parameterization of the eddy-flux term of (12'). That parameterization, and the corresponding development for the right-hand side of Equation (A.11), contains the real substance of the two theories and are dealt with in the review itself (Section 3).

## DISCUSSION

**A. DYUNIN:** The large clouds showing on your slide down-wind of the rock outcrops could be visually misleading. The main flow of snow will be invisible in the shadows. What is your opinion about this? Also, what is the basis for your use of a *constant* diffusion coefficient? I would like to thank Professor Radok for his stimulating contributions; much could be gained in the future through the merging of our two lines of investigation.

**U. RADOK:** Transport of snow below the 3 cm level is approximately equal in amount to that in the next 20 cm. Our highest measurements were made at 8 m, but we need data from much higher levels, even to 100 m. Drift extends much higher than is sometimes assumed.

**H. LISTER:** The deflation approach has no measure of the surface other than that manifest in the velocity, while the diffusion approach has a surface-roughness coefficient expressed in units of height. Surface characteristics can be very different, even with the same roughness coefficient, e.g. in dull weather, one day and five days after a snowfall. Drift on each of these days can be very different. Can you therefore suggest modifications of the roughness coefficient to incorporate these cohesion differences?

RADOK: The snow surface creates its own roughness.

P. M. B. FÖHN: The diffusion theory seems to yield only ratios of snow-drift densities; thus to gain the absolute drift density for a given level it is necessary to work from data obtained from a reference level. Would it be possible to calculate the drift density at a higher level by first deriving the "reference" drift density close to the ground using the Siberian method ( $n_z \propto v_1^2$ )?

RADOK: The drift density is quite constant for different levels, so that this is a small problem.