Doctoral Thesis

Wind Tunnel Investigations of Boundary Layer Conditions Before and During Snow Drift

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Wind Tunnel Investigations of Boundary Layer Conditions Before and During Snow Drift

A dissertation submitted to the
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presented by
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Abstract

Measurements of snow drift and wind velocity profiles were made over natural snow surfaces in a wind tunnel. Understanding the link between snow drift and the boundary layer is crucial for avalanche forecasting work and climate modelling. This work was carried out at the Swiss Federal Institute for Snow and Avalanche Research SLF in Davos, Switzerland.

The velocity boundary layer over snow follows the usual log-law relationship and can be characterised using friction velocity \( u^* \) and roughness length \( z_0 \). A definition of the threshold friction velocity \( u_{st} \) for snow drift has been developed based on single-height measurements of drift activity, and was applied to Snow Particle Counter (SPC) data using regression analysis. Measurements of drift activity, wind profiles and surface snow characteristics were made for 15 snow surfaces over 2 winters. \( u_{st} \) varies from 0.28 to 0.71 m s\(^{-1}\), and was influenced more by particle size than by ambient temperature or humidity. The threshold algorithm of Schmidt (1980) and a modified version as used in SNOWPACK agree well with the observed \( u_{st} \) if a small inter-particle bond size is assumed. Using hydraulic diameters obtained from images of surface snow crystals, the threshold parameter \( A \) is 0.18, about twice that for incohesive particles of the same diameter (Bagnold, 1941). Drifting snow particle sizes correlate at \( R^2 = 0.51 \) with surface particle sizes, and are similar to those seen in literature from field experiments.

Roughness lengths vary between snow covers, but are constant until the start of drift. This suggests that snow surfaces are hydraulically rough, and no indication was seen of a smooth hydraulic regime. The threshold \( z_0 \) varied from 0.04 to 0.14 mm, which is \( 0.021u_{st}^2/(2g) \), as seen for other aeolian drift (Owen, 1964). \( z_0 \) in the wind tunnel was lower than literature for field measurements, which shows the influence of macroscopic roughness and surface topology in the field. Small-scale roughness, probably from localised settling and clustering of grains on the snow surface, meant that correlations between particle size and roughness developed for sand, under-predict snow \( z_0 \).

Velocity profiles were also measured over open-celled foams. The foams used had permeability in the range \( 6 - 160 \times 10^{-9} \) m\(^2\); permeability of new snow is in the range \( 0.1 - 10^{-9} \) m\(^2\) (Albert and Schultz, 2002). The local pressure gradient was zero. Roughness lengths of the foam increased markedly with permeability, and also slightly with \( u_\ast \), in contrast to snow. \( z_0 \) correlates well with permeability. Velocity profile zero-plane displacement was observed in the more permeable foams, and increased with \( u_\ast \). Increasing \( u_\ast \) also increased \( z_0 \). Changing the foam depth from 6 to \( 38 \times \) the mean pore size had no obvious impact on either the displacement height or the roughness length. The zero-plane displacement for the foams with similar permeability to snow, was zero, even at \( u_\ast = 0.5 \) m s\(^{-1}\). These results suggest that flow within snow surfaces is not due to shear-driven momentum transfer from the atmospheric boundary layer to a flat snow cover.

The Lagrangian saltation model of Doerschot and Lehning (2002) was modified to deliver mass flux profiles by discretising particle trajectories. Particle trajectory initial conditions were parametrised using particle sizes, \( u_\ast \), and surface density, to allow use in a range of conditions. A particle size distribution was used on the surface. The simulated mass flux profiles are dominated by the near-ground flux, in contrast with field literature which shows less vertical variation. This difference is from superposition of different transport modes; in field measurements, small particles are partly convected by turbulence and moved further from the surface than by saltation alone, changing the mass flux profile.

Several wind tunnel experiments were simulated where the majority of mass transport was by particles large enough to be in saltation. Simulation results agree well with SPC-measured transport rates.

Further measurements are proposed into the role of fetch and height on the observations made here. Time-resolved drift and boundary layer velocity measurements are suggested, as current analysis has used time-averaged results. Further comparison with drift models requires knowledge of the trajectories in the wind tunnel, which would show the relative contribution of saltation and suspension.
Kurzfassung


Die Geschwindigkeitsgrenzschicht über Schnee zeigt die übliche logarithmische Höhenabhängigkeit und kann durch die Parameter Schubspannungsgeschwindigkeit $u^*$ und Rauhigkeitslänge $z_0$ beschrieben werden. Zur Bestimmung der Grenzschubspannungsgeschwindigkeit $u^*$ für Verfrachtung wurde eine Definition entwickelt, die auf einer Punktmessung der Driftaktivität aufbaut. Die Definition wurde benutzt, um mittels Regression Messungen der Drift im Windkanal mit dem Schneepartikelzähler (SPC) zu interpretieren. Driftaktivität, Windprofile und Schneeoberflächeneigenschaften wurden über den Zeitraum von zwei Wintern und für insgesamt 15 Schneedecken gemessen. $u^*$ variiert zwischen 0.28 and 0.71 m s$^{-1}$ und ist mehr von der Korngröße des Schnees als der Umgebungstemperatur oder -feuchte abhängig. Die Grenzwertformulierung von Schmidt (1980) und eine modifizierte Version wie sie in SNOWPACK verwendet wird zeigen eine gute übereinstimmung mit den Messungen wenn kleine Verbindungsbrücken zwischen den Schneekristallen (Bonds) angenommen werden. Wenn wir den hydraulischen Durchmesser, der durch Bildanalyse von Oberflächenkristallen bestimmt wurde, heranziehen, erhalten wir den Grenzparameter $A_0 = 0.18$, der in etwa doppelt so gross ist, als für kohäsionlose Partikel gleicher Grösse (Bagnold, 1941). Die Grösse der Partikel in Verfrachtung ist mit der Grösse der Oberflächenpartikel korreliert ($R^2 = 0.51$) und entspricht Literaturwerten aus Feldmessungen.

Die Rauigkeitslängen variieren stark für verschiedene Schneedecken, sind aber unabhängig von der Windgeschwindigkeit bis zum Einsetzen der Verfrachtung. Daraus folgt, dass Schneeoberflächen im Allgemeinen hydraulisch rau sind. Es wurde kein Hinweis auf ein hydraulisch glattes Regime gefunden. Die Rauigkeit beim Grenzwert für Verfrachtung bewegte sich zwischen 0.04 and 0.14 m, was 0.021$m^2/(2g)$ entspricht und mit Beobachtungen für den Windtransport anderer Materialien übereinstimmt (Owen, 1964). Die Rauigkeit im Windkanal ist deutlich kleiner als sie in der Literatur für Feldmessungen angegeben wird, was den Einfluss von makroskopischer Rauigkeit und Oberflächenformen im Feld zeigt. Auf der anderen Seite ist die Rauigkeit grösser als für flache Sandoberflächen angegeben wird. Während die Korngrösse die Rauigkeit flacher Sandoberflächen bestimmt, scheinen beim Schnee lokale Setzungsmechanismen und Koagulationen zu erhöhten Rauigkeiten zu führen.

Windprofile wurden auch über offenzelligen Schäumen als Analogon zu Schnee gemessen. Die Schäume hatten Durchlässkeitsbeiwerte von $6 \text{ bis } 160 \times 10^{-9} \text{m}^2$. Die Durchlässkeitsbeiwerte von Schnee reichen von 0.1 bis $10 \times 10^{-9} \text{m}^2$ (Albert and Schultz, 2002). Der lokale Druckgradient im Windkanal wurde für die Messungen genau auf Null eingestellt. Die Rauigkeit der Schäume nahm stark mit der Durchlässigkeit zu und zeigte im Gegensatz zum Schnee auch einen leichten Anstieg mit $u^*$. Eine Nullpunktsverschiebung des Windprofils konnte für die stärker porösen Schäume beobachtet werden, die ausserdem mit $z_0$ korrelierte. Die Dicke des Schaums, die in den Experimenten zwischen 6 und 38 mal der mittleren Porengrösse variierte, hatte keinen messbaren Einfluss auf Rauigkeit oder Nullpunktsverschiebung. Die Nullpunktsverschiebung für Schäume mit ähnlichen Durchlässigkeitsbeiwerten wie Schnee war null, sogar für die relativ hohe $u^* = 0.5$ m s$^{-1}$. Diese Resultate lassen folgern, dass Strömung in der Schneematrix nicht durch die Schubspannung über einer flachen Schneedecke zustande kommen kann.

Das Lagrange'sche Saltationsmodell von Doerschot and Lehning (2002) wurde erweitert, um Massenflussprofile rechnen zu können. Die Anfangsbedingungen für die Trajektorienberechnungen wurden...
Kurzfassung

eu als Funktion von Korngrösse, $u_*$ und Oberflächendichte beschrieben, um eine allgemeine Anwendbarkeit für verschiedene Situationen zu gewährleisten. Die Oberfläche wurde durch eine Korngrößenvverteilung beschrieben. Die berechneten Massenflussprofile haben ein Maximum nahe an der Oberfläche, während Feldmessungen oft eine gleichmässigere Verteilung zeigen. Der Unterschied sollte mit der überlagerung verschiedener Transportmodi zu tun haben: Kleine Partikel können bereits in Suspension sein, was durch das Modell nicht beschrieben wird und das Massenflussprofil glättet. Für die Berechnung von Windkanalexperimenten mit grossen Partikeln - so dass Saltation als der dominierende Transportprozess angenommen werden - stimmen Modell und Beobachtungen gut überein.

Zum Abschluss werden weitere Messungen vorgeschlagen, insbesondere um die Rolle der Beobachtungshöhe und die Entwicklung des Gleichgewichts im Windkanal zu untersuchen. Da die vorliegenden Ergebnisse auf zeit-gemittelten Daten beruhen, sollte auch mit zeitlich hochaufgelösten Messungen gearbeitet werden. Eine weitergehende Validierung von Verfrachtungsmodellen sollte sich auf die Messung von Trajektorien konzentrieren, weil somit auch zwischen den Transportarten Saltation und Suspension unterschieden werden kann.
Chapter 1

Introduction

Do nothing in haste; look well to each step; and from the beginning think what may be the end.

Edward Whymper, Scrambles amongst the Alps (1864)

Drifting snow has interested physical scientists, be they engineers, meteorologists or hydrologists for more than a hundred years. Initial scientific interest was prompted by the expansion of transport links into Alpine regions and High Latitudes, and the need to understand how those routes could be defended against drift. Arctic and Antarctic expeditions in the late 19th and early 20th centuries revealed the abrasive nature of drifting snow on people and landscape. In the latter half of the 20th century, expansion of leisure activities into Alpine regions, and the contribution that drifting snow makes to avalanche danger, have triggered renewed interest. Infrastructure development in remote or mountainous regions has increased the need for modelling drift transport, as no observations exist which can be used to predict the scale of the problem. Since the idea of climate change was mooted, the influence of snow and the snow surface on the atmosphere, and vice-versa, has also become an important research issue.

As forecasting avalanche hazard, predicting water budgets and predicting climate change impacts are predominantly computer-model based, there is a need for hard numbers to support these modelling activities. For a meteorologist, the requirements are simply put; how does snow differ (if at all) to other natural surfaces like grassland, trees or desert? In avalanche forecasting, the question is, how is snow moved by the wind, and how do deposition patterns change? For modellers, the question then becomes, how can this be used in a model of atmospheric circulation or snow metamorphosis?

Wind transports snow around the landscape, causing deposits that differ considerably from those that would be seen without wind. Snow is eroded from one location and deposited elsewhere in processes that are driven by regional weather systems but modified by local conditions, even down to the sub-meter scale. As Figure 1.1 demonstrates, the effects of wind are often complex in the real landscape. The foreground shows sastrugi, caused by erosion. The middle ground shows a ridge where snow has been deposited around its base, and a zone of flow separation is suggested at its crest. The pock-marked slope in the background shows influences of erosion and deposition. That these effects are all seen in a region of some $30 \times 30$ m gives some idea of the small-scale variability which occurs. Simulating the wind field around this landscape alone would merit another doctoral thesis, and results obtained would only apply in this particular case. Instead, the aim of this work is to investigate the behaviour of a flat snow surface as wind blows across it. This will be done under controlled conditions, using a wind tunnel. Of particular interest are the relationship between the wind at a given measurement height and the wind or forces at the surface, the wind speed at which surface snow starts to move in the wind (drift) and the links between these data and the snow itself. Snow characteristics will be limited to single grain and bulk descriptions, such as particle size, shape and overall snow density.

The remainder of this chapter describes how wind and snow interact, and describes in more detail the goals of the thesis, which was carried out using the wind tunnel at the Swiss Federal Institute for Snow and Avalanche Research SLF, in Davos, Graubünden, Switzerland.
Chapter 1. INTRODUCTION

Figure 1.1: Wind-sculpted snow above St. Antönien, Graubünden, Switzerland.

1.1 The atmosphere and the snow pack

The development of the snow pack with time and in space is strongly influenced by the local climate. As snow is a high-temperature material, existing at temperatures very close to its melting point, the material from which it is formed is found in all 3 phases simultaneously within the snowpack. The matrix of the snow pack is formed of ice, while the pore space is occupied with air, saturated with water vapour. It is not uncommon to find free water, particularly near the surface during spring, while a thin adsorbed water layer can be found on the surface of ice particles at all temperatures. This existence at a high temperature makes snow sensitive to energy fluxes. Energy loss from the snow pack causes the snow to cool and hence vapour and liquid accretes on to the snow, while energy fluxes into the snow increase the snow temperature and cause increased sublimation and melting.

The snow pack acts as an insulating cover on the ground. The thermodynamic thickness of a covering is described by the Biot Number ($Bi$). This is defined as the ratio of heat transfer by convection at the surface (sometimes called advection) to heat transfer by conduction through the material;

\[
Bi = \frac{\text{convection}}{\text{conduction}} = \frac{\alpha h_{ss}}{k_e} \quad (1.1)
\]

where $\alpha$ is the surface convective heat transfer coefficient, $h_{ss}$ the thickness of the snow cover and $k_e$ the effective thermal conductivity. In the case of a 1 m thick snow cover, where $\alpha$ is $O(100)$ W m$^{-2}$ K$^{-1}$ and $k_e$ is $O(0.1)$ W m$^{-1}$ K$^{-1}$ (Sturm et al. 1997), this gives $Bi \approx 1000$. The heat transfer in insulating covers where $Bi > 1$ is dominated by convection, rather than conduction of heat. This means that non-linear temperature gradients develop within the snow, and so the processes of mass transfer, and hence snow metamorphosis, occur at different rates at different depths within the snow.

Energy fluxes in a snow pack are to, and from, the atmosphere, from solar radiation and a geothermal flux from the ground. The energy fluxes at the surface can be defined by the energy balance

\[
k_e \frac{dT}{dz} = q_{sh} + q_h + q_{lw} + q'' \quad (1.3)
\]
where $T$ is the temperature at a depth $z$, $q_{sh}$ is the sensible heat flux, $q_{lh}$ is the latent heat flux, $q_{LW}$ is the net long-wave radiation (incoming - outgoing) and $q''$ the energy flux from precipitation. Short wave radiation from the sun is strictly a source term within the snow pack, as snow is translucent at short wavelengths. Energy is supplied to the base of the snow pack by the geothermal heat flux. In Alpine environments, the net energy flux at the surface is $O(100) \ W \ m^{-2}$, and shows strong diurnal and seasonal variation. The geothermal flux is $O(0.10) \ W \ m^{-2}$, and is constant at depths of over 50 m. Hence the process of metamorphosis is dominated by the conditions at the atmosphere-snow interface.

The sensible and latent heat fluxes $q_{sh}$ and $q_{lh}$ are driven by turbulent exchange of heat and moisture at the snow surface. This turbulent exchange is strongly influenced by the wind speed and temperature and humidity gradients in the air. The exchange can be defined as functions of the surface roughness length for momentum, $z_0$, and the friction velocity $u^*$; two parameters which define the wind velocity profile above a surface.

The shear stresses acting in the boundary layer are also important. Above a certain minimum friction velocity ($u^*_t$), the surface becomes mobile, and drift starts. Drift erodes material from the snow pack, and deposits it elsewhere in the landscape. Alternatively, the eroded snow can be suspended in the atmosphere for long enough that it can sublime, and the mass is lost (in the short term) to the snow surface. Both saltating and suspended material alters the aerodynamic characteristics of the surface, changing the surface energy balance as well.

Measurements for other drifting material, such as sand and soil, show that above the drifting material, $z_0 \propto u^*^2$ [Owen, 1964]. Measurements also show links between $u^*_t$, $z_0$ and the particle dimensions which have not been investigated for snow.

### 1.2 Wind tunnel measurements

Linking field measurements of surface processes over snow to the snow itself are inherently difficult. Firstly, results obtained over long periods of time must be reduced to useful mean values, such as $u^*_t$, $z_0$, and transport rates. Unpredictability of wind speeds and directions means that data has to be filtered for direction and speed, and also means that local characterisation of snow types and surface properties may be irrelevant, if snow is being transported in from elsewhere. Secondly, wind speeds and thus aerodynamic properties of the surface are commonly measured at 2-10 meters above the surface in boundary-layer meteorological studies. Therefore, the aerodynamic properties of at least the same, if not much larger, length of surface are being convected to the instrument, and so topography within the measurement footprint might be expected to be more important than the snow characteristics. The only locations where these influences are less pronounced are the flat, smooth and uniform slopes of the Arctic, Antarctica and Greenland, where katabatic (and thus strong and uniform) winds dominate. However, running experiments in these locations is expensive.

Compared to trying to understand the wind field in Figure 1.1 and the resulting drift, a wind tunnel with a snow-covered floor has several basic advantages. Wind is constrained to flow down the major axis of the tunnel. Turbulence and wind speed are controlled by the flow-conditioning, and the exhaust fan, not the weather. Snow is of known heritage, and can be measured and sampled with ease. The inlet and exit conditions are known, and the velocity boundary layer over the snow can be measured.

The Swiss Federal Institute for Snow and Avalanche Research SLF in Davos, Switzerland, operates the only wind tunnel worldwide designed for use with natural fresh snow. The tunnel is some 13 metres long, and has 1 m × 1 m cross-section. The flow is conditioned to approximately match Outside conditions. Boundary layer profiles and drifting snow are measured over a 4 m long snow cover.

### 1.3 Drift models

Drift models have been developed to allow numerical simulation of drift. These range from the correlation-based models, considering simply wind speed at a certain height above ground, to models which resolve multiple snow flakes moving in turbulent flows. Models can be broadly divided into Eulerian simulations and Lagrangian simulations. In Eulerian simulations, drifting snow is simply another property that is convected by the wind field, such as temperature or moisture. Snow transport is then modelled by diffusion and convection, and requires the diffusion coefficient for snow in air to
be known. Alternatively, Lagrangian simulations can be used, where individual particle’s tracks are resolved as they are moved by the wind, with or without resolution of turbulence. A knowledge of the particle size and number in motion is required to estimate the total horizontal mass flux. Eulerian models are generally computationally cheap, depending on the flow resolution which is required. By comparison, Lagrangian models have a higher computational requirement, as particle trajectories have to be calculated through the air. If feedback on the flow is included, the computational load increases further in both cases.

Models of snow drift are essentially identical to other models of aeolian transport, until processes specific to snow are introduced. These snow-specific models could include such effects as sublimation, surface erosion or hardening, entrainment rates and surface rebound.

Previous work at SLF has led to the development of the semi-Lagrangian model of Doorschot and Lehning (2002), which calculates the mass flux from near-ground snow drift (saltation). Aspects of this model will be compared to results from the wind tunnel.

1.4 Focus of this thesis

This thesis concentrates on the form of the boundary layer with and without drift. Previous research has shown that the atmospheric velocity boundary layer over snow is often logarithmic, and hence the roughness length \( z_0 \) and friction velocity \( u_* \) can be used to characterise the boundary layer (Budd et al., 1966). Correlations between feature lengths in the snow and \( z_0 \) without drift are investigated, and the measured threshold compared to predictions for sand and snow. The modification of the boundary layer by drift is investigated, and measurements are compared to saltation simulations.

The questions that this study was designed to address are;

1. Is \( z_0 \) independent of \( u_* \), until drift starts?
2. What length scale in the snow best correlates with \( z_0 \) before drift?
3. What is the threshold \( u_* \) required for drift, and how does it vary with different snow types?
4. When drift occurs, does the boundary layer over drifting snow follow the hypothesis of Owen (1964), so that \( z_0 \propto u_*^2 \)?
5. Do we observe a similar mass flux and particle size at a given height to that predicted using the model of Doorschot and Lehning (2002)?

This thesis describes the development of the SLF wind tunnel and analytical methods to answer these questions.

In Chapter 2, atmospheric velocity boundary layer theory is introduced. The velocity profile parameters \( z_0 \) and \( u_* \) are defined, and the influence of drift discussed. Several algorithms for the prediction of the drift threshold are introduced.

Chapter 3 details the SLF wind tunnel and the instrumentation used. It describes the experiments that were performed, and considers uncertainty and repeatability. Uncertainty and repeatability are shown to be sufficient to resolve influences of the surface and drift on the boundary layer. Drift is demonstrated to be in equilibrium when free-stream wind speeds are kept constant.

Chapter 4 considers the onset of drift in detail, as this is a major focus of current snow pack and boundary layer models. Drift thresholds are compared to predictions from algorithms, and reasons for discrepancies identified. The roughness length with and without drift is given, and compared to predictions for other granular media.

Chapter 5 looks at the microstructure of snow and compares it to commercially available, porous foam. Boundary layer measurements are made over this foam, and results compared to those for snow. This shows the influence of material length scales on \( z_0 \) and how \( z_0 \) varies with \( u_* \) without drift.

Chapter 6 takes an existing saltation model and compares simulation results to measurements in the wind tunnel. Mass flux profiles as a function of height and friction velocity are simulated, and compared to measurements and assumptions made in the previous chapters.

The summary and conclusions in Chapter 7 summarise the results from the previous chapters and discuss their implications. Finally, the last section highlights some avenues for future research.
Chapter 2

Theory

This chapter introduces basic atmospheric velocity boundary layer theory, concentrating on the steady-state velocity profile. The log-law profile is introduced and the two key parameters $u_*$ and $z_0$ are defined. The cause of snow drift, and the influence of drifting snow on velocity boundary layers is discussed. Numerical modeling of snow transport is also addressed.

2.1 Velocity boundary layer

The time-averaged, turbulent, neutrally-stratified, velocity boundary layer over an immobile rough surface is described by the log law profile. In this profile, the velocity $u(z)$ at a height $z$ above the surface is a function of the surface roughness length, $z_0$ and the friction velocity, $u_*$, so that

$$u(z) = \frac{u_*}{\kappa} \ln \left(\frac{z}{z_0}\right)$$  \hspace{1cm} (2.1)

where $\kappa$ is the von Karman constant, taken here as 0.41 (for a review of experimental results in literature, see p. 289 of Garratt, 1994).

Many field weather stations are only single level, i.e.; the wind speed is only measured at a single height, much greater than $z_0$. Calculating $u_*$ then requires an estimate of $z_0$. If measurements are made at two or more heights, a linear fit to measurements of $u(z) = f(\ln(z))$ can be used to calculate the $z$-intercept, where $u(z) = 0$ and hence $z = z_0$; the friction velocity is given by $u_* = \kappa d\ln(f(\ln(z)))$. The result from a multiple-height profile depends on the validity of the log-law profile, the accuracy of the velocity and height measurements, and the value of $\kappa$. The effect of $z_0$ and $u_*$ on the mean velocity profile is shown in Figure 2.1.

Flow in a wind tunnel is constrained to have net flow in only one direction. This is the $x$ direction, and $x$ increases in the direction of flow. The $y$ direction is parallel to the floor, but normal to $x$. $z$ is the distance from the surface of the tunnel. Because the flow is turbulent, the velocity measured in the direction $x$, at a height $z$ ($u(z)$), can be defined using Reynolds' decomposition as the sum of a mean and turbulent fluctuation $u(z) = u(z) + u(z)'$, where the time-averaged mean velocity is $\bar{u}(z)$ and the turbulent fluctuation is $u(z)'$. The same is true of the flow in $y$ and $z$, where the velocities are $v(z) = \bar{v}(z) + v(z)'$ and $w(z) = \bar{w}(z) + w(z)'$ respectively. The mean velocities normal to the tunnel axis should be zero.

The effect of the small fluctuations in the fluid velocity ($u'$, $v'$ and $w'$) are to transport momentum. This gives rise to localised Reynolds stresses in the fluid, where the mean stress $\tau$ in a plane at a particular location is given by the product of the velocity fluctuations and the density. In the $xz$ plane this would be

$$\tau_{xz} = -\rho u'w'.$$ \hspace{1cm} (2.2)

Averaged over a sufficient period of time, the mean value of the Reynolds stresses in the $xy$ and $yz$ planes when flow is homogenous and non-accelerating, is zero. For simplicity, this thesis uses $\tau = \tau_{xz}$.

The total shear $\tau$ in a boundary layer is the sum of the viscous stresses and the Reynolds stresses in the flow, and is given by
The contribution of the Reynolds stresses and the viscous stresses varies with height above the surface. The height above the wall can be defined in terms of the wall layer thickness, $\delta_v$, where $\delta_v = \nu / u_*$.

In the wind tunnel at $-10^\circ$C and 835 hPa, $\nu$ is approximately $1.48 \times 10^{-5}$ m$^2$ s$^{-1}$. The typical $u_*$ is $0.3$ m s$^{-1}$. This gives a wall layer thickness $\delta_v = 50 \mu m$. Below $z = 5\delta_v$, in the viscous sublayer, the viscous stresses dominate. Above $z = 30\delta_v$, shear stresses are dominated by the Reynolds stresses, and in this region Equation 2.1 applies. In a boundary layer flow, this region extends to about $0.2 \times$ the boundary layer thickness. The total stresses are assumed to be approximately constant in the logarithmic region, which is also called the constant stress layer in boundary-layer meteorology.

In the constant shear layer, the mean shear stress $\tau$ can also be defined as a function of $u_*$ and the air density $\rho$, and is given by

$$\tau = \mu \frac{du}{dz} + \rho u' w'. \quad (2.3)$$

and so $u_*$ is equivalent to $\sqrt{u' w'}$.

Measurements in the constant shear layer can be used to calculate $u_*$ and $z_0$ by fitting velocity profiles to the log-law. However, that approach requires that the constant-shear region is correctly identified. One option is to deliberately measure the velocity at points outside of the estimated constant-shear region, and use these to identify the part of the velocity profile that follows the log-law. This is practical in a wind tunnel, where the bottom of the log region is at $z \approx 1.5$ mm, and the boundary layer thickness can be $O(1)$ m, and hence the log region could be expected to extend over $0.0015 < z < 0.2$ m. In the free atmosphere, the boundary layer can be $O(1000)$ m thick, and therefore the constant shear layer may extend to 200 m, which precludes this approach. However, the depth of the constant shear layer does give more freedom for simply placing wind measurement devices at two or more heights on a 10 or 30 m tower. Alternatively, the eddy-correlation technique can be used in the constant shear layer, whereby $u'$ and $w'$ are measured directly, and $u_*$ can be determined using equation 2.4. This gives $u_*$ at the height of the measurements, which in the log region should also be independent of height.

$$\tau = \rho u_*^2, \quad (2.4)$$
2.1. VELOCITY BOUNDARY LAYER

2.1.1 Hydraulic regimes

Surfaces are generally divided into 3 hydraulic regimes, depending on the height to which wall roughness penetrates the different regions of the boundary layer, and the influence they have on the drag [Schlichting and Gersten, 2003, p. 522].

Smooth regime If the roughness elements of height \( d \) perpendicular to the wall are similar in size to the wall layer thickness \( \delta_c \) and so \( \delta_c/d \approx 1 \), the elements are submerged in the viscous sub-layer of the boundary layer. In this case, the roughness elements do not affect the wall drag and the wall is considered hydraulically smooth. The limit to this regime is \( d/\delta_c = 5 \).

Transitional regime As \( d/\delta_c \) increases, the roughness elements project out of the viscous sub-layer but do not reach the boundary with the overlap layer of the boundary layer until \( z = 70\delta_c \). This is known as the transitional flow regime and occurs while \( 5 \leq d/\delta_c \leq 70 \); viscous forces still influence the drag, and hence \( z_0 \) is a function of \( u_* \).

Rough regime As \( d/\delta_c \) increases further, and elements penetrate into the overlap layer, the roughness elements directly influence the wall drag. The drag is independent of the flow Reynolds number, and hence \( z_0 \) (which is an expression of the surface drag coefficient) is independent of \( u_* \). This is known as the hydraulically rough regime, and so fully rough flow requires that surface roughness elements have a size \( d \geq 70\delta_c \).

Schlichting and Gersten (2003, p. 530) give \( z_0 = d/25 \) in the fully rough regime when the wall is covered with densely packed sand grains over a solid surface. Combining this with the particle size requirement gives the rough regime as \( z_0/\delta_c \geq 70/25 \). As \( \delta_c = \nu/v_* \), a roughness Reynolds number can be defined, \( Re_* = z_0u_*/\nu \), in which case the rough regime occurs when \( Re_* > 2.8 \). By comparison, Greeley and Iversen (1985) give \( d/30 \leq z_0 \leq d/8 \), depending on the distribution of sand on the surface, and hence the start of the rough regime is in the range \( Re_* = 2.33 \) to \( 8.75 \) (see Table 2.2 for details), and Andreas (1987) and Bintanja and van den Broeke (1995) assume the rough region is \( Re_* > 2.5 \). Given the uncertainty about the onset of the fully rough regime in terms of \( Re_* \), it seems more sensible to stay with a definition of the rough regime as the flow regime where \( z_0 \neq f(u_*) \), implying that \( d \geq 70\delta_c \).

Measurements over snow and ice by Bintanja and van den Broeke (1995) and Andreas et al. (2004) suggest that at low friction velocities \( (u_* < 0.2 \text{ m s}^{-1}) \), \( z_0 \) decreases as \( u_* \) decreases, rather than increasing as expected from smooth regime theory. This decrease may be because \( z_0 \) is ill-defined at low shears, or that the necessary measurement accuracy cannot be achieved. For \( u_* \geq 0.2 \text{ ms}^{-1} \), measurements of both Bintanja and van den Broeke (1995) and Andreas et al. (2004) show a slight increase in \( z_0 \) with \( u_* \), although no information is given regarding the presence of drift.

The presence of a smooth regime may be linked to flow within the snow surface, a phenomena known as ventilation. If there is some displacement of flow into the surface, it may be possible for large particles or agglomerations to protrude into the overlap region of the flow. This would allow the surface to set the drag even at low \( Re_* \). Under wind tunnel conditions, a rough regime would therefore require projection of obstacles about 3.5 mm into the flow. It is important to note that not all elements would have to reach the same height; the roughness elements can be sparse, for example trees on a grassy plane. However, these elements can define the roughness length of the whole plane, when observations are made at heights significantly above the height of the elements (Garratt, 1994).

2.1.2 Roughness lengths over natural surfaces

Natural surfaces are often considered to be rough surfaces. Roughness lengths are then said to be independent of the friction velocity. Data collected for boundary layers over uniform natural surfaces shows a large variation in \( z_0 \), ranging from 0.001 m for soils up to \( \approx 5 \text{ m} \) for tropical forests. Measurements over large areas of flat snow give \( z_0 \) lower than other natural surfaces, except in mountainous regions. Results in literature are summarised in Table 2.1.

The roughness length \( z_0 \) can be considered a length scale for the boundary layer. If snow is considered a granular medium, the individual grains are the most obvious features on the surface, and several length scales are apparent, including the diameter, pore size and center-to-center distance. Because of
Table 2.1: Roughness lengths for various surfaces from literature. Measurements over snow surfaces are highlighted.

<table>
<thead>
<tr>
<th>$z_0$ (m)</th>
<th>Description</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>4-5 × 10$^{-1}$</td>
<td>Rocky Mountains (USA)</td>
<td>Stull (1988)</td>
</tr>
<tr>
<td>0.3-3 × 10$^{-1}$</td>
<td>small hills, 100-180 m elevation</td>
<td>Stull (1988)</td>
</tr>
<tr>
<td>9.2 × 10$^{-1}$</td>
<td>pine forest, 15.8 m trees</td>
<td>Garratt (1994)</td>
</tr>
<tr>
<td>4 × 10$^{-1}$</td>
<td>savannah, 8 m trees</td>
<td>Garratt (1994)</td>
</tr>
<tr>
<td>3.2 × 10$^{-1}$</td>
<td>pine forest, 12.4 m trees</td>
<td>Garratt (1994)</td>
</tr>
<tr>
<td>2.3 × 10$^{-2}$</td>
<td>thick grass, 0.1 m high</td>
<td>Garratt (1994)</td>
</tr>
<tr>
<td>4.3 × 10$^{-3}$</td>
<td>‘flat snow’ on the Seward Glacier, Canada</td>
<td>Lougeay (1970)</td>
</tr>
<tr>
<td>2 × 10$^{-3}$</td>
<td>sparse grass, 0.015 m high</td>
<td>Garratt (1994)</td>
</tr>
<tr>
<td>1-20 × 10$^{-3}$</td>
<td>snow in mountainous terrain</td>
<td>Doorschot et al. (2004)</td>
</tr>
<tr>
<td>1-10 × 10$^{-3}$</td>
<td>soils</td>
<td>Garratt (1994)</td>
</tr>
<tr>
<td>1 × 10$^{-4}$</td>
<td>snow surface on Filchner-Ronne Ice Shelf, Antarctica</td>
<td>Heinemann (1989)</td>
</tr>
<tr>
<td>3 × 10$^{-6}$</td>
<td>Antarctic blue ice</td>
<td>Bintanja and van den Broeke (1995)</td>
</tr>
</tbody>
</table>

† Precise details and references are in Appendix 4, Table A6 of Garratt (1994).

these different scales, drag over a granular surface has been found to be a function of both the size and shape of the grains and the spacing between the individual grains that form the surface (Morris, 1955; Wooding et al., 1973; Greeley and Iversen, 1985; Schmeekle and Nelson, 2003; Schlichting and Gersten, 2003). Attempts have been made to parametrise the roughness length of surfaces based on length scale correlations. These are summarised in Table 2.2.

Table 2.2: Roughness length parameterizations from literature.

<table>
<thead>
<tr>
<th>Surface Conditions or usage</th>
<th>$z_0$</th>
</tr>
</thead>
<tbody>
<tr>
<td>sand (Schlichting and Gersten, 2003, p. 530)</td>
<td>$d/25$</td>
</tr>
<tr>
<td>- close packed</td>
<td>$d/30$</td>
</tr>
<tr>
<td>nearly spherical elements' (Greeley and Iversen, 1985)</td>
<td>$d/8$</td>
</tr>
<tr>
<td>- close packed</td>
<td>$d/30$</td>
</tr>
<tr>
<td>- center-to-center distance $\approx 2d$</td>
<td>$d/10$</td>
</tr>
<tr>
<td>- center-to-center distance $\gg 2d$</td>
<td>$d/10$</td>
</tr>
<tr>
<td>snow</td>
<td>$0.003 + d/10$</td>
</tr>
<tr>
<td>- weather station data interpretation (Lehning et al. 2000a)</td>
<td>$d/10$</td>
</tr>
<tr>
<td>- saltation modelling (Doorschot and Lehning, 2002)</td>
<td>$d/10$</td>
</tr>
</tbody>
</table>

2.1.3 Effects of permeable surfaces

Snow is a permeable medium. Other permeable surfaces in nature include ground cover such as forestry, arable crops or grassland growing over impermeable soil. The velocity boundary layer over a permeable surface is usually altered to allow for penetration of flow into the surface by including a zero-plane displacement height $z_d$ into the normal log-law velocity profile (Stearns, 1970; Arya, 2001), so that

$$u(z) = \frac{u_*}{\kappa} \ln \left( \frac{z + z_d}{z_0} \right).$$

(2.5)
This results in a non-zero or ‘slip’ velocity at the upper surface of the porous media, which is taken as \( z = 0 \). The zero-plane displacement height \( z_d \) can either be assessed from a fit to profile and shear measurements, or assessed from local feature heights, \( h_0 \). For small vegetation it is assumed that \( z \gg z_d \) and so \( z_d \) can be neglected, while in larger vegetation, \( z_d/h_0 \approx 2/3 \) (Garratt [1994]), although this varies with the canopy density. In snow it is rarely clear where an impermeable surface is; on snow on ice caps, it might be assumed that it is the surface of multi-year ice, although it might just as easily be an ice layer within the snow pack. The same is true in fresh snow in temperate climates; without a detailed manual profile of the snow, it is difficult or impossible to identify the depth at which snow becomes impermeable (Colbeck, 1991). As total snow heights are often only 1-2 m, and permeable layers might only form a few centimeters at the surface, the net displacement might then be only a few millimeters. When analysing the data from weather stations over snow, the displacement height is often ignored as data is obtained at 1 m or more above the surface and so it is assumed that \( z + z_d \approx z \). Data from Kobayashi (1969) and simulations from Albert and Schultz (2002) suggest that the displacement could be non-zero, and so should be considered when measuring near to the snow surface (when \( z \approx z_d \)) but these studies were not able to isolate the influence of shear driven flow from pressure driven flow, or assess the influence of the snow permeability.

Length scales in porous media include the pore size, dimensions of the elements forming the matrix, and the material depth. In materials composed of irregular objects, the dimensions become less well defined, and can be converted to surface or volume-averaged values. A common measure is the specific surface area \( SSA \), defined as the surface area of the solid per total volume. This can be converted to a representative diameter \( d_v \) if the porosity \( \phi \), defined as the void fraction, is known (e.g. Breugem et al., 2006), so that

\[
d_v = 6(1 - \phi)/SSA.
\]  

2.2 Snow drift

Drift is a generic term used to describe mass transfer by erosion of discrete particles from a granular or loose surface and subsequent deposition. Drift is usually considered to occur in any or all of 3 different modes simultaneously, shown schematically in Figure 2.2. These modes are distinguished between by the motion of the particles.

The mode which occurs at lowest wind speeds is known as reptation or creep, and is simply the rolling of material along the surface, with no movement into the boundary layer. Consequently any influence on the boundary layer is limited, and seen only as an increase in roughness \( z_0 \) compared to the immobile surface. Initially this is the only mode of mass transfer during ‘drift’, but as wind speeds and hence free-stream shear \( \tau \) increase, gradually contributes less to the total mass transport than other modes (Nemoto and Nishimura, 2004).

The next mode, saltation, occurs when particles bounce over the surface in parabolic paths, easily visible in photographs, reaching a height of up to 0.2 m above the surface. Within the layer formed by the saltating particles, there is significant momentum transfer from the wind to particles. This causes an observable modification to the wind profile, first measured in saltating sand by Bagnold (1941) and since seen in saltating snow by Maeno et al. (1979, Figure 4). The boundary layer above the saltating media sees the saltation as increased roughness (see section 2.3).

![Figure 2.2: Different modes of snow transport over a snow surface](image-url)
The final mode is known as suspension, where particles are convected along with the eddy structures of the turbulent flow, and hence effectively behave as flow tracers. This results in mass being transferred further from the surface than would be the case with saltation. Studies have shown that depending on particle sizes and wind speeds, a considerable proportion of the total mass transported by drifting material may be moved in this way (Bagnold, 1941; Budd et al., 1966; Mann et al., 2000; Nishimura and Nemoto, 2005).

The tendency of particles to enter into suspension is a function of the ratio $u_*/u_f$, where $u_f$ is the particle settling velocity. The settling velocity of a spherical particle in still air, assuming Stokes flow, is given by

$$u_f = \frac{\rho_p g d^2}{18 \mu},$$

where $\rho_p$ is the particle density and $\mu$ the dynamic viscosity of air. For typical wind tunnel conditions, $\mu = 1.64 \times 10^{-5} \text{ kg m s}^{-1}$. When $u_*$ becomes greater than $u_f$, particle trajectories start to be influenced by turbulence (Nalpanis, 1985). At a typical $u_* = 0.3 \text{ m s}^{-1}$, this means that ice particles smaller than 99 $\mu$m diameter would be expected to show signs of modified saltation. At $u_* > 10u_f$, particles behave purely as tracers and enter into suspension. At $u_* = 0.3 \text{ m s}^{-1}$, this would require particles of 31 micron diameter or smaller.

Suspended particles have been observed to reach heights up to 1 km above the ground (Mahesh et al., 2003), while modelling suggests that they can cross similar distances before returning to the surface (Déry and Tremblay, 2004); this has even been suggested as a means of seeding remote Mediterranean islands with miniature snails (Kirchner et al., 1997). Because the momentum required to transfer particles into suspension has already been provided at some distance upstream, or by saltation, it is usually presumed that a classical log-law velocity profile can re-establish itself and that flows with suspension can be treated as mixed, two-phase flows (Déry and Yau, 2001; Déry and Tremblay, 2004). This is supported by velocity profiles shown in Nemoto and Nishimura (2004, Figure 11), where drifting snow causes a noticeable reduction in velocity near the ground (in the saltation layer) but less so above 0.1 m, despite the presence of drift. There is also some indication that the two-phase flow should include modified turbulent parameters, particularly the turbulence intensity (Bosse, 2005). This approximation as a mixed flow should hold so long as the material in suspension does not alter the thermodynamics of the flow locally, for example by suspended snow subliming in strong sunlight or low relative humidity, and so causing density changes through the boundary layer.

The split into different modes of drift, with mass interchange between them, is the basis for most modelling of mass transfer by snow drift (Pomeroy et al., 1993; Liston and Sturm, 1998; Gauer, 1999). The process is split into a saltation model and a separate suspension model, which share an interface some distance above the ground, across which a particle or mass flux is defined. Creep is often simply neglected in modelling as having negligible contribution.

Because most natural surfaces include a range of particle sizes, and wind speeds and hence shear are not constant, these processes can occur simultaneously over a surface. This leads to transport to a greater height than would be expected by saltation alone (e.g., Nemoto and Nishimura, 2004).

### 2.2.1 The drift threshold

Particles begin to drift when the forces acting on them are sufficient to separate them from the bed, and lift them above the surface. The onset of drift is described as the threshold, and when no other particles are drifting, is due purely to aerodynamic forces.

A review of all forces acting on a drifting particle is given in Gauer (1999). At the threshold, the most important forces are the fluid forces, body forces and inter-particle forces. A sketch of the forces acting on a single particle on a horizontal, granular surface, raised slightly above its neighbours, is shown in Figure 2.3. This can be considered as the simplest realistic case; while a single particle bonded to a smooth surface might be the simplest analytical case, it is not a particularly useful analogy for snow surfaces.

The moment equilibrium about the downstream pivot point, assuming that the particle is only bonded at one point, is given by

$$Da + Lb + M = Wb + Ic,$$  \hspace{1cm} (2.8)
where \( D \) is the drag force, \( a \) is the height that the particle rises above its upstream neighbour, \( L \) the lift force, \( b \) is the horizontal distance from the centre of mass to the pivot, \( M \) the moment of the particle, \( W \) the particle mass, and \( c \) is the radial distance from the centre of mass to the pivot.

The lift and drag acting on the particle can be determined from the shear acting on the particle using the coefficients of lift and drag (%(\text{lift} \ C_L \text{ and } \text{drag} \ C_D \text{ respectively}), while the moment is a function of the shear, size, volume and moment coefficient (%(\text{moment} \ C_M \text{ of the particle);} 

\[
\begin{align*}
D &= C_D \rho u^* a^2 d^2 \\
L &= C_L \rho u^* b^2 d^2 \\
M &= C_M \rho u^* c^3 
\end{align*}
\] (2.9)

Including the effect of buoyancy and now denoting the density of the solid as \( \rho_p \), substituting Equation (2.9) into Equation (2.8) gives

\[
C_D \rho u^* a^2 d^2 + C_L \rho u^* b^2 d^2 b + C_M \rho u^* c^3 = (\rho_p - \rho) g \frac{d}{2} \left( \frac{d}{2} \right)^3 b + Ic. \] (2.10)

By assuming that the distances \( a, b \) and \( c \) are some fraction of \( d \), and hence \( a = a_1 d, b = b_1 d \) and \( c = c_1 d \), and that the actual weight of the particle is somewhat less than a solid particle of the same size, where the weight is given by \( \eta_p W \), we come to

\[
\rho u^* d^3 (C_D a_1 + C_L b_1 + C_M) = \frac{1}{6} (\rho_p - \rho) \eta_p g \pi d^4 b_1 + d Ic_1, \] (2.11)

and hence at the threshold,

\[
u^*_t = \sqrt{\frac{\frac{1}{6} (\rho_p - \rho) \eta_p g \pi d^4 b_1 + d Ic_1}{\rho(C_D a_1 + C_L b_1 + C_M)}}. \] (2.12)

Noting that the density \( \rho \) is \( \rho_{\text{air}} \) and that \( \rho_p \) is the density of ice, \( \rho_{\text{ice}} \), and assuming that the cohesive force \( I \) is zero allows this to be simplified to simplification;

\[
u^*_t = A \sqrt{\frac{\rho_{\text{ice}} - \rho_{\text{air}}}{\rho_{\text{air}} g d}}, \] (2.13)

where \( A \) is the threshold parameter, and includes the geometrical factors of the material and its arrangement with respect to its neighbours;

Figure 2.3: Fluid and body forces acting on an individual particle on a porous surface, built up from individual grains of a material (after Greeley and Iversen, 1985). Forces shown are lift \( L \), drag \( D \), moment \( M \) about the downstream point of contact which forms the pivot point, a binding force \( I \) due to a bond, and weight \( W \).
Alternatively, if the cohesion is not zero, the threshold parameter becomes a function of the particle form and aerodynamics, and the forces between the particles:

\[
A_c = A \sqrt{1 + \frac{6}{\pi g} \frac{c_1}{\eta p b_1} \frac{I}{(\rho_{\text{ice}} - \rho_{\text{air}}) d^3}} \tag{2.15}
\]

where \(A_c\) indicates the threshold parameter for the cohesive case. The effect of cohesion is to scale \(A\), allowing Equation (2.13) to be used for both the cohesive and non-cohesive cases. Importantly, the binding force \(I\) might also be expected to be a function of the bond area and thus particle size, reducing the influence of \(d\) on the ratio \(A_c/A\) which is implied by Equation (2.15).

### 2.2.2 Predicting the drift threshold

Equation (2.13) can be used to predict the drift threshold of a granular material if \(A\), the particle size and density, and the fluid density are known. An exhaustive series of experiments reported by Bagnold (1941, p. 101) showed that the threshold parameter, \(A\), was about 0.1 for sand larger than 0.2 mm, in air. Other experiments, summarised in Greeley and Iversen (1985), show that this is generally true for large particles, not just sand. The same experiments also show that for particles smaller than 0.2 mm, the cohesive forces from electrostatic forces, water, sintering and chemical bonding become important.

For particles where the particle friction Reynolds number \(Re = \rho d u_{\text{sat}}/\mu\) is less than 5, the value of \(A\) rises from \(\approx 0.1\) to \(0.2\) at \(Re_{\text{sat}} \approx 1\).

When there is already significant drift occurring, the movement of particles into the flow might be due to impact of drifting material on the bed. Because of this, Bagnold suggested that the threshold should be described in terms of an entrainment threshold, where drift is caused solely by aerodynamic forces, and a rebound threshold, where drift is sustained by already-moving material. Measurements by Maeno et al. (1979) indicate that \(u_{\text{sat}}\) for snow with increasing wind speeds are about 1.2 times those measured during decreasing wind speeds.

Although snow is porous and appears granular and so similar to sand, there are many differences between the two materials. The actual area of a snow crystal which is exposed to wind forces could be larger than that for a sphere of similar extent, because the particle might be porous (see Figure 5.1). The actual area of a snow crystal which is exposed to wind forces could be larger than that for a sphere of similar extent, because the particle might be porous (see Figure 5.1). The actual area of a snow crystal which is exposed to wind forces could be larger than that for a sphere of similar extent, because the particle might be porous (see Figure 5.1). The actual area of a snow crystal which is exposed to wind forces could be larger than that for a sphere of similar extent, because the particle might be porous (see Figure 5.1). 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Compared to Equation (2.13), Equation (2.16) is more complex and requires more detailed information, particularly about the snow form. The snow form information is expressed in terms of particle diameter \( d \), bond diameter \( d_b \), sphericity \( SP \) and the 3-dimensional co-ordination number \( N_3 \).

In this thesis, the grain diameter \( d \) is determined from image processing of digitised snow particle photomicrographs. The photomicrographs have a resolution 5 to 10 \( \mu m \). The snow particle outlines are detected by thresholding of the images. The mean diameter of a particle is taken as the mean distance through its centroid to diametrically opposing points every 2\(^\circ\) around the circumference. A simple numerical mean of this mean diameter for a sample is used for the sample average, \( d_I \). The bond diameter \( d_b \) can be obtained from 3-dimensional imaging of the snow structure, but this approach is difficult and slow. The sphericity \( SP \) is also measured from images, and then mapped to the range [0, 1] using the definitions of Lesaffre et al. (1998) to give \( SP \). Particle perimeter is also measured from the images, using a cubic-spline fit to the outlines, which may be a weak function of the resolution and particle form. The area is calculated as the zeroth-order moment of the outline (eqn. 2.2 in Seul et al., 2000).

In predictive modelling, \( d \), \( d_b \) and \( SP \) could be obtained from snow cover modelling (e.g. SNOWPACK; Bartelt and Lehning, 2002).

The 3-dimensional coordination number \( N_3 \) can be measured from 3 dimensional snow samples or calculated from a correlation given in Lehning et al. (2002a) which is based on data in Edens and Brown (1991). The correlation uses the bulk density of the snow, \( \rho_s \), to calculate the mean co-ordination number:

\[
N_3 = 1.43 - 7.56 \times 10^{-5} \rho_s + 5.15 \times 10^{-5} \rho_s^2 - 1.73 \times 10^{-7} \rho_s^3 + 1.81 \times 10^{-10} \rho_s^4.
\]

Equation (2.17) gives \( N_3 = 1.53 \) for snow of density 50 kg m\(^{-3}\), increasing to 1.78 at 100 kg m\(^{-3}\). This approach is used in SNOWPACK (Bartelt and Lehning, 2002).

Alternatively, the weather conditions can be used to predict the threshold. For example, Li and Pomeroy (1997) gave a correlation between the threshold 10 m wind speed and air temperature. Based on a fit to observations in the Canadian prairies, they proposed that:

\[
U_t(10) = 9.43 + 0.18T_a + 0.0033T_a^2
\]

where \( T_a \) is the air temperature (\(^\circ\)C) measured at 2 m above the surface. They reasoned that the air temperature controlled or heavily influenced most factors relevant to the initial motion of snow, such as the grain shapes, sizes and bonding to neighbouring particles, and hence no other predictive variable was necessary.

### 2.3 Influence of drift on the boundary layer

Transport of material by a fluid boundary layer requires transfer of momentum from the fluid to the material. This causes a deceleration of the fluid in the drifting layer when the free-stream velocity remains constant. In the case of air and sand where the density of the sand is high compared to the fluid, this causes a clear increase in roughness when seen from above the saltating region, and was conclusively demonstrated by Bagnold (1941, pp. 57-76) and Owen (1964). Owen’s hypothesis was that the roughness over a drifting surface should increase compared to the stable surface, such that

\[
\frac{u(z)}{u_*} = \frac{1}{\kappa} \ln \left( \frac{2gz}{u_*^2} \right) + D' \tag{2.19}
\]

where \( D' \) is a constant. Equating this to Equation (2.1), it is seen that this gives

\[
z_0 = C \frac{u_*^2}{2g}, \tag{2.20}
\]

where the constant \( C \) is \( \exp(-D' \kappa) \). Owen explained this increased roughness as the particles on the surface being raised into the boundary layer during drift. From conservation of energy, if particles
leave the surface with a vertical speed proportional to \( u_\ast \) \(^2\) (Nishimura and Nemoto, 2005), the height the particles reach would be proportional to \( u_\ast \) \(^2\).

From a survey of literature data, Owen found \( D' \) to be 9.7, i.e. \( C = 0.021 \), over sand and soil. A literature survey by Tabler (1980) gave \( 0.011 \leq C \leq 0.028 \) for smooth snow covers with long fetch, then König (1985) found \( C = 0.012 \) over snow-covered ice, while Bintanja and van den Broeke (1995) measured \( C = 0.032 \) under similar conditions. Other factors also influence the roughness length, raising it from that solely due to drift; Liston and Sturm (1998) used \( C = 0.12 \) to account for ‘topographically variable’ terrain. This high value of \( C \) serves to raise \( z_0 \) at all \( u_\ast \) to account for the topography, and add a small increase due to drift, effectively treating drift as a second-order roughness effect compared to the landscape (see Table 2.1). Using an arbitrarily high \( C \) implies that the roughness length of the topology is a function of \( u_\ast \), and so a better approach would be to take the maximum value of \( z_0 \) for the terrain, and that due to drift, at a given \( u_\ast \).

At the onset of saltation, Equations (2.20) and (2.1) must both apply. Therefore, \( z_0 \) before drift (if it is constant) can then be used to calculate \( u_{st} \) from Equation (2.20). If \( u_\ast \) then increases beyond \( u_{st} \), the minimum \( z_0 \) which is seen by measurements above the saltation layer is given by Equation (2.20). Owen’s relationship gives the minimum \( z_0 \) because topographic influences could increase the apparent \( z_0 \) at a given \( u_{st} \), as was observed by Doorschot et al. (2004).

Figure 2.4 demonstrates the attractiveness of Equation (2.20); usually, single level measurements of wind speed do not allow \( u_\ast \) to be estimated, as \( u_\ast \) and \( z_0 \) are coupled. If \( z_0 \) is known, or the form of \( z_0 = f(u_\ast) \) is known and monotonic, as in Equation (2.20), then \( u_\ast \) can be estimated from \( u(z) \). This helps with interpreting meteorological measurements from masts, for example, but also in more practical situations, such as estimating wind speeds at which drifting (or suspended) snow might obscure visibility on roads or runways.

![Figure 2.4](image-url)

Figure 2.4: The conclusion of Owen’s First Hypothesis. The bold black line is equation (2.20). The contours are 10 m wind speeds calculated from the log-law profile, (Equation (2.1)), showing the possible combinations of \( u_\ast \) and \( z_0 \) which give the same wind speed.

Owen’s second hypothesis (also in Owen, 1964) offers an insight into the nature of drift. He notes that once the drift threshold is reached and the free-stream friction velocity \( u_\ast \) is equal to \( u_{st} \), a surface begins to release particles. Increasing \( u_\ast \) such that \( u_\ast > u_{st} \) causes increased entrainment, and the increased momentum transfer to the drifting particles from the air causes the friction velocity at the surface to drop to \( u_{st} \). This suggests two important points;
Drift is sustained: If $u_s > u_{st}$, drift should continue so long as the surface remains unchanged.

**Entrainment rates increase with** $u_s$: The mass flux and the number of particles in the air increase as the friction velocity increases. Once the boundary layer has achieved equilibrium, the surface friction velocity and thus shear is constant, the lift and drag which act at the surface are also constant and so the size of particles being entrained should not change. During gusts, the boundary layer is not in equilibrium, increasing $u_s$ at the surface, and so there is an increased chance that larger particles could be entrained.

This gives a potential method for defining the drift threshold. Because drift is sustained, if $u_s$ is increased step-wise, the threshold should be clearly visible from a plot of either the mass flux or particle count of drifting snow against $u_s$. A simple regression can then be used to find the threshold.

### 2.4 Modelling snow transport

Snow transport models try to model the transport of snow particles by the wind. An overview of modelling methods and the rational behind the choices is given in [Hutter (2005)](https://journal.ofhutter.com), and is summarised here.

Some models take a correlation-based approach to the mass transported by wind (e.g., Pomeroy and Gray, 1990; Sørensen, 1991, which are both based around $u_s$ and $u_{st}$). While these models are computationally efficient, they are limited in the physics of the processes that they resolve.

Direct Lagrangian simulation of all particle paths, surface collisions and other processes is extremely costly computationally (e.g., Boss [2005]). Because of this, models tend to be split between the two main mass transport regimes, saltation and suspension, with each treated separately. This split is shown diagrammatically in Figure 2.5.

**Figure 2.5:** Modelling approaches used for snow drift. Left, processes of mass transport from the surface and in the boundary layer. Right, the numerical modelling methods used in each region.

Field data, for example from [Nishimura and Nemoto (2005)](https://journal.ofnishimura.com), shows that the two modes of mass transport are clearly visible when horizontal mass flux is measured as a function of height. The maximum mass transport is found in the lowest 0.05 m of the boundary layer, and as $u_s$ increases, the mass transport above 0.1 m increases much more than that near to the ground. These two regions demonstrate the split between near-ground saltation and suspension further from the ground.

Many saltation models take a Lagrangian approach and directly resolve the paths that the particles follow over the surface (e.g., Nalpanis, 1985; Anderson and Haff, 1988; McEwan, 1993; Doorschot and...
Lehning, 2002; Nemoto and Nishimura, 2004). A finite number of particles are entrained from the surface, and then the trajectories that they follow are determined by the interaction between the boundary layer and particles. The equations which describe the motion of the particle in a steady-state cross-wind are then:

\[
\frac{d^2x}{dt^2} = -0.75 \frac{\rho_{air}}{\rho_{ice}} \frac{u_r}{d_p} C_D \left( \frac{dx}{dt} - u(z) \right)
\]

(2.21)

for motion parallel to the ground (direction \(x\)), where \(C_D\) is the drag coefficient, \(u_r\) the relative velocity between particle and local wind field, and \(dt\) denotes differentiation with respect to time. In the vertical direction, the equation is

\[
\frac{d^2z}{dt^2} = -0.75 \frac{\rho_{air}}{\rho_{ice}} \frac{u_r}{d_p} C_D \frac{dz}{dt} - g.
\]

(2.22)

Determining suitable values of \(C_D\) has been the subject of many papers (amongst others Chepil, 1958; Sumer, 1984; Mollinger and Nieuwstadt, 1996; le Roux, 2004), and for simplicity’s sake, it is often simply assumed that drifting snow particles are spherical ice grains.

A Lagrangian saltation model might include momentum exchange with the surrounding fluid, not just simple one-way (fluid to particle) coupling. The form of the boundary layer might be based on a simple time-average approach, such as Doorschot and Lehning (2002), include a parametrisation of turbulence (Nalpanis, 1985) or be determined by LES simulation (Nemoto and Nishimura, 2004), which can then also be used to drive the particle motion (Bosse, 2005). For a more complete discussion of saltation modelling, see Doorschot (2002).

Suspension models consider snow concentration as a bulk property of air which is simply convected through space in a similar way to other fluid properties. This is described as the Eulerian approach. A diffusion scheme is used, based on the fall velocity of the particles, and effects such as sublimation can also be included in the simulation. This approach is summarised in various papers from S. Déry and R. Bintanja (Déry and Taylor, 1996; Déry and Yau, 2001; Déry and Tremblay, 2004; Bintanja, 2000a, b, 2001a), and is similar to simulation methods used for sand in air, or sediment transport in water. The major requirement of a snow suspension model in comparison to other aeolian suspension models is the inclusion of a sublimation model, which allows for mass transfer to and from the drifting snow to the surrounding atmosphere.

The split between saltation and suspension models results in some problems. Chief among these is how to model situations where both saltation and suspension occur. The usual approach is to apply the results of a saltation simulation as the lower boundary condition for the suspension model, either as a number of particles or as a concentration of snow per volume of air (Gauer, 1999; Déry and Yau, 2001), and allow the suspension model to develop with this lower boundary condition. Unfortunately, this approach neglects the fact that towards the top of the saltating region, the contribution to the total horizontal mass from suspension can already be significant. That is to say, the assumption of a single height at which mass transport switches from saltation to suspension is somewhat simplistic, and best suited to conditions where only the integral mass transport is of interest. It is therefore better to choose a situation where the snow transport is predominantly by saltation (i.e. large particles, low \(u_*\), measurements less than 0.1 m from the surface) or by suspension (i.e small particles, high \(u_*\), measurements more than 0.10 m from the snow surface) and model the transport using the corresponding physical processes.

2.5 Summary

In this chapter, boundary layer theory has been introduced with reference to the earth’s surface. Two main areas, firstly the relationship between roughness lengths, \(u_*\) and length scales in snow, and secondly the drift threshold and its relationship to surface conditions, have already been identified as areas where useful work can be done. Existing knowledge has been summarised in these areas. Some uncertainty remains about the form of the boundary layer over snow because of the presence or otherwise of a hydraulically rough regime. Also, its permeability may result in a displacement depth, and this issue will be investigated as part of this thesis. The influence that aeolian drift has on the boundary layer has also been described, particularly the dependency of \(z_0\) on \(u_*^2\) once drift starts. The physics of
drifting snow have also been discussed, concentrating on the implications that this might have for modelling snow transport. Because the two main modes of above-ground transport are inherently different in terms of the physics that governs the particle motion, simulation of mass transport by drift must wait until the details of particle sizes and $u_*$ in the wind tunnel are known, at which time a suitable suspension or saltation model can be chosen.
Chapter 3

Experiment setup and measurements

We slaved until we were nearly dead-beat, and then we found something else to do until we were quite dead-beat.

A. Cherry-Garrard, *The worst journey in the world* (1922)

In this chapter measurements of drifting snow in the SLF wind tunnel are introduced. Measurement techniques and equipment are detailed. A typical experiment in the wind tunnel is described, and the data is compared to literature test cases.

3.1 The SLF wind tunnel

The SLF wind tunnel is situated at 1650 m a.s.l in the Flüelatal, above Davos, Graubünden, Switzerland. The tunnel itself is housed in a converted Swiss Army bunker, and snow falls are collected outside the bunker in trays. These trays are then moved intact into the bunker with as little disturbance as possible, allowing the behaviour of a natural snow pack to be investigated. Trays are used for experiments when they contain between 0.06 and 0.16 m depth of undisturbed snow. Tests are also possible using sieved snow or other granular or porous materials.

The SLF wind tunnel is 14 m long, with a $1 \times 1$ m cross section. The tunnel operates in suction, drawing air from outside the bunker through a honeycomb, nets and a 4:1 contraction into the tunnel. Downstream of the contraction are spires designed to create large scale vortices. These are followed by regular roughness elements for 4 m and then 4 m of fine-fibre, long pile carpet. This gives a logarithmic boundary layer of $\approx 0.25$ m depth. This is followed by a 4 m test section, where the bottom is replaced with the snow trays. The height of the snow trays is adjusted so that the top of the snow coincides with the uppermost fibres of the carpet. The trays are open in the streamwise direction ($x$), allowing an uninterrupted boundary layer development. Center-line wind speeds up to $20 \text{ m s}^{-1}$ can be reached. A sketch of the tunnel is shown in Figure 3.1.

![Figure 3.1: Schematic view of the SLF wind tunnel.](image)

The wind tunnel is a new facility at SLF. It was relocated from EPFL in Lausanne to Davos in 2002, and was initially commissioned by Dr. Jean-Daniel Rüedi and Thomas Exner. The author started work at SLF in winter 2003/2004. An important theme of the work presented here is to demonstrate that the
wind tunnel flow field and the behaviour of the drift are similar to test-cases from literature, and that stable, repeatable operation is possible.

3.2 Flow characteristics

The wind tunnel flow has been set up to mimic flow in the atmospheric boundary layer. This work was predominantly carried out by D. Ambühl during Summer 2004 (Ambühl 2004). The aim of that work was to establish a logarithmic boundary layer at the inlet to the test section, using a combination of turbulence generators and roughness elements. A wooden floor was used in the test section. Velocity profiles were measured using a hot-wire anemometer mounted on a traverse mechanism, which translated horizontally (\(y\)-direction) and vertically (\(z\)-direction). The mean velocity contours and turbulence intensity of the flow at one free-stream condition are shown in Figure 3.2.

![Figure 3.2: Example velocity and turbulence profiles at the test section inlet. Velocities are measured by hot-wire anemometer. (a) Mean velocity contours. (b) Turbulence intensity. Data from Ambühl (2004)](image-url)

The flow velocity is proportional to the log of the measurement height. The logarithmic region is approximately 0.2 m high. \(u_*\) and \(z_0\) calculated from measurements at different spanwise locations are summarised in Table 3.1. Variation in measured \(u_*\) across the tunnel is less than 10%, while variation in \(z_0\) is more than 40%. Inlet roughness lengths are the same order of magnitude to those that would be expected over snow, so adjustment to the different roughness over snow should be rapid. The increase in \(z_0\) on one side of the tunnel, compared to the other, is a result of a leak into the tunnel which decelerates the flow. The leak was caused by the traverse mechanism. This result highlights the sensitivity of \(z_0\) to small changes in the flow, or measurement uncertainty.

Based on these results, velocity profile measurements were made along the center-line of the wind tunnel, in the region \(0.25 \leq y \leq 0.5\) m. Data used to calculate \(u_*\) and \(z_0\) were generally limited to \(z \leq 0.2\) m or less, and so should not be affected by the deceleration caused by the leakage.

Although the hot wire system delivers detailed, high-frequency velocity measurements, it is not suitable for measurements with drifting snow. A vertical profile requires several minutes to complete, which increases the potential for formation of a surface crust or other snow metamorphosis. Drifting snow particles can also impact the wire, either breaking small wires (1 micron diameter), or causing spikes in the bridge voltage as the wire temperature recovers, or causing system instabilities. These events must be filtered out to obtain usable data. Because of these problems, velocity measurements were made with more rugged devices, which are described in the following section.
3.3 Method

The primary aim of the SLF wind tunnel is to allow measurements over a snow surface. The location and setup of the tunnel is tailored to this goal. A chief requirement of experiments over a snow surface is that the air temperature over the snow be similar to that found outside the facility. The bunker housing the tunnel is shaded from the sun by mountains from December to March and is not heated. To minimise temperature differences between the tunnel, the bunker and the surroundings, the building doors were opened, and the wind tunnel run, before experiments started. Experiments then started when the temperature difference between the outside and the test-section was less than 2 K. During experiments, test-section inlet temperatures were found to vary by less than 0.5 K from the ambient temperature. The snow surface temperature was measured by thermocouple before being transferred into the tunnel, and was within 2 K of the ambient temperature.

Most of the experiments over snow surfaces described here were carried out using fresh snow which was deposited by precipitation. This means that the number of experiments during any one winter is limited by the number of storms. An example of two snow packs laid down by precipitation is shown in Figure 3.3. One set of trays are smooth and uniform, while the other, deposited during a storm, are unusable; the significant changes in height in both spanwise and streamwise directions would result in significant accelerations and flow inhomogeneity.

Snow can also be sieved into the trays from snow which has built up around the building. When sieved snow was used, the surface of the snow was levelled using a wooden edge, dragged along the surface. Sieved snow is always marked in this document, as it might have different geometric and

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Table 3.1: Fits to the test-section inlet velocity profiles shown in Figure 3.2(a). Data is fitted using the log-law (Equation 2.1) to profiles at several different spanwise ($y$) positions at the same fetch. Data used to find $u_*$ and $z_0$ is limited to $z \leq 0.2$ m. $R^2$ is the value of the coefficient of determination for the fit.

<table>
<thead>
<tr>
<th>$y$ (m)</th>
<th>$u_*$ (m s$^{-1}$)</th>
<th>$z_0$ (mm)</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.25</td>
<td>0.67</td>
<td>0.12</td>
<td>0.97</td>
</tr>
<tr>
<td>0.35</td>
<td>0.70</td>
<td>0.16</td>
<td>0.98</td>
</tr>
<tr>
<td>0.45</td>
<td>0.71</td>
<td>0.16</td>
<td>0.98</td>
</tr>
<tr>
<td>0.55</td>
<td>0.68</td>
<td>0.14</td>
<td>0.98</td>
</tr>
<tr>
<td>0.65</td>
<td>0.78</td>
<td>0.26</td>
<td>0.99</td>
</tr>
<tr>
<td>0.75</td>
<td>0.82</td>
<td>0.32</td>
<td>0.99</td>
</tr>
<tr>
<td>mean</td>
<td>0.73</td>
<td>0.19</td>
<td></td>
</tr>
<tr>
<td>$\sigma$</td>
<td>0.06</td>
<td>0.08</td>
<td></td>
</tr>
</tbody>
</table>
physical characteristics to snow which was deposited by precipitation, and the surface is likely to be flatter than a purely ‘natural’ snow pack. The effort involved with preparing sieved snow samples meant that this method was only used on a few occasions.

The trays were then moved into the wind tunnel, and the height was adjusted so that the snow surface was level with the upper fibres of the carpet in the upstream region. An array of anemometers were then positioned with a traverse mechanism at the maximum possible distance downstream from the start of the snow pack (3 m). Anemometers were positioned at nominal heights of 50, 100 and 150 mm above the snow surface, and exact heights were measured with a ruler to the highest point of the surface vertically below. Velocity profile measurements allowed the validity of the log-law to be checked, and $u^*_*$ and $z_0$ were calculated by a least-squares fit to Equation (2.1). The snow particle counter (SPC) was positioned 50 mm above the snow at the same fetch as the anemometers. Figure 3.6 gives a view into the tunnel, showing the equipment positioned above the snow surface.

Experiments were also carried out with artificial porous media on the floor of the test section (see Section 5). In these experiments the upper surface of the medium was adjusted to coincide with the top of the carpet. The SPC was left out of the tunnel.

The tunnel speed is then increased stepwise from $\approx 3 \text{ ms}^{-1}$ in the free-stream to $\approx 15 \text{ ms}^{-1}$. These intervals are described as measurement ‘plateau’ and usually last from 3-4 minutes. Longer intervals cause significant erosion of a loose surface, while shorter intervals do not allow the system to reach equilibrium.

The logarithmic wind profile is defined as a time-averaged profile, not at an instant, and meteorological observations frequently use average values over $10^3$ to $10^4$ s (Wolfe and Nickling, 1993). Because of this, instantaneous observations of snow drift and wind profiles or $u_*$ cannot be used, and data must also be time-averaged. Before time averaged data can be obtained, the measurement plateau are identified manually from the data. An example of the selected velocity plateaux and the variation of the velocity in the plateaux is shown in Figure 3.4.

![Figure 3.4](image)

Figure 3.4: Raw wind speed and selected plateaux data during an experiment. Data is taken from measurements at two heights during a test over snow on 9th March 2005. Every 10th velocity measurement is shown. Lower plots show the velocity distribution at each height, during the plateau.

Figure 3.4 shows that the velocity fluctuations in the boundary layer at 50 and 100 mm above the snow surface were less than 10 % of the mean value. This was measured using propellor anemometers, which have a response time of less than 1 second and act as a low-pass filter, removing higher frequency changes in wind speed due to turbulence. Other measurements showed that fluctuations of the free-stream velocity, as measured by pitot-static tube at the test section inlet during the plateau, were less than 1% of the mean. This low free-stream fluctuation, and the lack of skew or distortion of
the velocities measured near the ground suggests that the wind tunnel was operating stably and not beating or chasing, and so no filtering of the time series during the plateaux was carried out.

During an experiment, the following data were measured:

Reference wind speed. A dynamic pressure was obtained from a pitot-static tube at the inlet to the test section and measured using an MKS Baratron Model 220D capacitance pressure sensor, connected to an MKS PR400 control unit. Pressure readings were obtained via an RS232 interface, and quoted pressure accuracy is 0.12% of the reading. This gives a velocity accuracy of 0.06%.

Wind profile. Two methods were used to measure the velocity profile. The first was a continuous measurement at two heights, measured by anemometer. The second was a dynamic pressure profile measuring at 11 heights through the boundary layer.

Continuous profile. The velocity profile at the mid span after 3 m fetch was measured at 50 and 100 mm above the surface by Schiltknecht MiniAir Micro anemometer. Quoted velocity accuracy is 0.5% of the full-scale ($20 \text{ m} \cdot \text{s}^{-1}$), and 1.5% of the reading. These also measure temperature to $\pm 0.5^\circ\text{C}$. Voltage output signals ($0...20 \text{ m} \cdot \text{s}^{-1} = 0...1 \text{ V}$, $-20...140^\circ\text{C} = 0...1.6 \text{ V}$) were obtained via an Agilent 34410A multimeter with GPIB connection to a National Instruments PCI interface.

Dynamic pressure profile. The MiniAirs were supplemented during winter 2005/2006 with a total pressure rake. The fetch to the rake was 2.5 m, which was positioned at $y = 0.3 \text{ m}$. The rake had 10 pressure taps, arranged to give 3 points below the level of the SPC, an extra 3 or 4 points in the logarithmic region (depending on the height of the logarithmic region) and the remainder in the upper part of the boundary layer.

Measurements of dynamic pressure were made during the constant-velocity plateaux. Differential pressure measurements were made between the total pressure and a reference static pressure probe 50 mm away from the tunnel centre-line at the same fetch. Pressures were measured using an MKS 120D differential pressure transducer, connected via a Scanivalve to the rake pressure taps. This gives the dynamic pressure at each measurement height, $P_{\text{dyn}}$. The dynamic pressure at the rake was then converted to velocity assuming incompressible flow, and from Bernoulli’s Equation, it follows that

$$u = \sqrt{\frac{2P_{\text{dyn}}}{\rho_{\text{air}}}}$$

where $\rho_{\text{air}}$ is the air density inside the tunnel.

Temperature and humidity were measured by Rotronics ‘Hygroclip’ S3 solid-state relative humidity (RH) and temperature sensors. The chips are protected from snow accumulation by a wire-mesh shield. Quoted accuracy at $23^\circ\text{C}$ is $\pm 1 \%\text{RH}$ and $\pm 0.3 \text{ K}$. Sensors are calibrated relative to a water surface. Voltage output signals ($0...100 \%\text{RH} = 0...1 \text{ V}$, $-40...60^\circ\text{C} = 0...1 \text{ V}$) were obtained via the Agilent 34410A.

Tunnel pressure difference. The static pressure drop in the wind tunnel was measured as a differential pressure by a Validyne DP103 capacitance sensor with $\pm 350 \text{ Pa}$ range and $\pm 10 \text{ V}$ output signal, acquired through the Agilent 34410A. The positive side of the sensor was connected to the static pressure measurement of the pitot-static at the inlet to the test section. The negative side of the sensor was connected to a short tube, open to atmosphere.

Weather conditions. A weather station continually monitored absolute atmospheric pressure, temperature and humidity. Data was recorded at 1 minute intervals using a Campbell Scientific CR10X data logger, and retrieved at the end of experiments.

Drift. A Niigata Electric Snow Particle Counter ‘SPC-S7’ was used to detect drift after 3 m fetch, 50 mm above the snow surface. The SPC is a single-beam photelectric detector where particles interrupt a near-infrared light beam, causing a change in sensor voltage. (similar to devices described in

1 see [www.snowcon.com/seihin](http://www.snowcon.com/seihin)
The basic operating principle of the SPC is shown in Figure 3.5. The SPC measures drift in a detector plane 0.6 mm deep in the direction of flow, 2.5 mm high and 25 mm wide. A high frequency signal is processed by the system’s own data processing unit and converted into 1 Hz mass flux and particle size data. Data was logged via RS232 interface and the SPC system’s own software, giving the number of particles in 32 equivalent diameter classes in the approximate range $50 \leq d \leq 500 \mu m$.

After an experiment, data from the systems were synchronized. The air density in the tunnel was calculated from atmospheric pressure, the static pressure drop to the test section, and the inlet temperature using the equation of state. Tunnel pressure drop was $O(100) \text{ Pa}$; given an ambient pressure of around 835 hPa, this could be neglected at the cost of a 0.1% inaccuracy in the estimated air density.

![Figure 3.5: Principle of operation of the SPC. An infrared (IR) beam of depth $\delta x$ is projected across a gap to an IR detector. A snow particle passes through the beam, reducing the IR light received by the detectors. The drop in incident light is proportional to the projected area of the particle and thus $d_p^2$, and the particle velocity $u_p$ can be obtained from the transit time, where $u_p = (\delta x + d_p)/\delta t$.](image)

![Figure 3.6: Instrumentation set up inside the wind tunnel. Close up inside the test-section of (left) total pressure rake, (center) anemometer array and (right) SPC.](image)
3.4 VELOCITY PROFILES

Measurements of the velocity profile over several different surfaces were made using the dynamic pressure rake. Several examples are highlighted here. These confirmed that the logarithmic wind profile was present over these different surfaces, and that the MiniAir anemometers were correctly positioned in the logarithmic region.

3.4.1 Calculating \( u^* \) and \( z_0 \)

Friction velocity \( u^* \) and roughness length \( z_0 \) were calculated from the mean velocity profiles in each plateau using best fits to the log law (Equation 2.1). Neutral stratification of the boundary layer was assumed over all surfaces. Fits were constrained to those regions where \( u \) showed a linear relationship to \( \ln(z) \).

In cases where only 2 measurement heights were available, \( z_d \) was set to zero to prevent numerical instabilities. This automatically gives a coefficient of determination \( R^2 = 1.0 \). In flow over porous media, the zero-plane displacement depth \( z_d \) was not fixed a priori as zero. Instead, the regression routine used an iterative technique after Clauser (1954). Initial ranges were chosen for \( z_d \), 10 values of \( z_d \) were chosen from that range, and the data was then fitted using (2.5). This gave \( R^2 \) for the fit, and the range of \( z_d \) was adjusted to bracket the best-fitting data. This process was repeated until the value of \( z_d \) which gave the highest \( R^2 \) stabilised. If the solution converged on a value of \( z_d \) lower than the resolution of the instrument heights (1 mm), \( z_d \) was assumed to be negligible, and the fit to the velocity profile recalculated using (2.1).

3.4.2 Smooth solid surface

A series of measurements were made over a solid, smooth wooden surface in a low streamwise pressure gradient (\( |dP_w/dx| < 0.02P_{dyn} \), where \( P_w \) is the static pressure measured on the wall). These velocity profiles are shown in Figure 3.7. These (and other) velocity profiles include repeated measurements at \( z \approx 200 \) mm, where the repeated \( u(z) \) agree to \( \pm 0.15\% \). The kink in the velocity profiles at \( z \approx 50 \) mm suggests an internal boundary layer has formed. This implies that the solid wooden floor is markedly smoother than the carpet upstream (\( z_0 \approx 0.19 \) mm). Because of the internal boundary layer, local \( u^* \) and \( z_0 \) were calculated from a fit to the velocity profiles below the interface to the upper boundary layer. These are shown as filled points in Figure 3.7 and the fit is the solid line. Values are summarised in Table 3.2.

<table>
<thead>
<tr>
<th>Profile</th>
<th>( u^* ) m s(^{-1} )</th>
<th>( z_0 ) mm</th>
<th>( z_d ) mm</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.18</td>
<td>0.005</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>2</td>
<td>0.23</td>
<td>0.007</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>3</td>
<td>0.25</td>
<td>0.005</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>4</td>
<td>0.29</td>
<td>0.006</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>5</td>
<td>0.33</td>
<td>0.007</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>6</td>
<td>0.35</td>
<td>0.006</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>7</td>
<td>0.34</td>
<td>0.003</td>
<td>0.0</td>
<td>&gt;0.99</td>
</tr>
</tbody>
</table>

The boundary layer over a solid floor is a common test case. Literature values for the normalised velocity profile over a smooth wall (see e.g. Schlichting and Gersten 2003, or Kundu and Cohen 2004) are plotted with these measurements in Figure 3.8.

In the logarithmic region, \( u^+ = \frac{u}{u^*} = \frac{\ln(y^+)}{C_s} + C_s \), where \( C_s \) is a constant. The results in Figure 3.8 show that each velocity profile has at least 6 points in the logarithmic region, where \( u(z)/u^* \) is proportional to \( \ln(z + z_d) \). Schlichting and Gersten gives \( C_s = 5 \) for a smooth wall, which is plotted in Figure 3.8 while
Chapter 3. EXPERIMENT SETUP AND MEASUREMENTS

Figure 3.7: Velocity profiles over a solid, smooth wooden floor. Profiles were measured by dynamic pressure rake at several spanwise ($y$) positions at the same fetch. Data is fitted using the log-law (Equation 2.5), and the fit is limited to $z \leq 50$ mm. Different markers indicate different measurement plateaux. The calculated $u_*$ and $z_0$ are summarised in Table 3.2.

Figure 3.8: Normalised velocity profiles over a solid wall. The solid line is the ‘universal velocity profile’ for flow over a solid wall in zero pressure gradient (Schlichting and Gersten, 2003, p. 535-531). Markers correspond to those shown in Figure 3.7. Different boundary layer regions (viscous sublayer, buffer layer and logarithmic region) are identified between approximate ranges of $y^+$.

Bhaganagar et al. (2004) gives $C_s = 5.5$. The standard deviation of $z_0$ measured over the solid smooth wall, from the expected value for a smooth floor, is 22% if $C_s=5$, and 16% if $C_s=5.5$. 
3.4. VELOCITY PROFILES

3.4.3 Porous surface

A series of measurements was made over a porous foam with similar characteristics to snow (100 mm thick Regicell 30, see Section 5.1). The aim of these measurements was to demonstrate that a zero-plane displacement could be detected, if porosity was high enough. The velocity profiles shown in Figure 3.9(a) have a tendency towards an acceleration near the surface, suggesting a slip velocity and a measurable $z_d$. Velocity profiles corrected for the zero-plane displacement height are shown in Figure 3.9. Points used to calculate $u_*$, $z_0$ and $z_d$ are shown as filled points in Figure 3.9. Values are summarised in Table 3.3. The zero-plane displacement shows a gradual increase with $u_*$, as expected (Jackson, 1981).

![Figure 3.9: Velocity profiles over a porous foam floor. Profiles were measured by dynamic pressure rake. Data is fitted using the log-law (Equation 2.5). Different markers indicate different measurement plateau. The calculated $u_*$, $z_0$ and $z_d$ are summarised in Table 3.3.](image)

![Figure 3.9(b): Corrected velocity profiles for zero-plane displacement.](image)

Table 3.3: $u_*$ and $z_0$ from velocity profiles over a porous surface. These are from fits to the velocity profiles shown in Figure 3.9. Data is fitted using the log-law (Equation 2.5).

<table>
<thead>
<tr>
<th>Profile</th>
<th>$u_*$ (m s$^{-1}$)</th>
<th>$z_0$ (mm)</th>
<th>$z_d$ (mm)</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.37</td>
<td>0.87</td>
<td>2.9</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>2</td>
<td>0.47</td>
<td>1.07</td>
<td>3.3</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>3</td>
<td>0.56</td>
<td>1.24</td>
<td>3.7</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>4</td>
<td>0.65</td>
<td>1.47</td>
<td>4.3</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>5</td>
<td>0.72</td>
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<td>4.3</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>6</td>
<td>0.81</td>
<td>1.75</td>
<td>5.0</td>
<td>&gt;0.99</td>
</tr>
<tr>
<td>7</td>
<td>0.88</td>
<td>1.95</td>
<td>5.7</td>
<td>&gt;0.99</td>
</tr>
</tbody>
</table>

3.4.4 Snow surfaces

The velocity profiles over a snow surface were measured using shorter measurement intervals than over the wooden floor or porous media. This was because at high velocities, drift becomes significant, causing erosion and moving the rake in relation to the ground. Analysis of the dynamic pressure time series suggested that the number of measurements was sufficient to give an accurate mean value. Re-
results are plotted in Figure 3.10. The calculated $u_*$ and $z_0$ are summarised in Table 3.4. The data used to fit the log-law is limited to $z \geq 0.1$ m when drift is detected by the SPC.

![Velocity plots](image)

Figure 3.10: Velocity profiles over snow. Profiles were measured by dynamic pressure rake. Data is fitted using the log-law (Equation 2.5). Different markers indicate different measurement plateaux. The calculated $u_*$, $z_0$ and $z_d$ are summarised in Table 3.4.

Table 3.4: $u_*$ and $z_0$ from velocity profiles over snow. These are from fits to the velocity profiles shown in Figure 3.10. Data is fitted using the log-law (Equation 2.5). $P(\text{drift})$ is the frequency of drift.

<table>
<thead>
<tr>
<th>Profile</th>
<th>$u_*$ (m s$^{-1}$)</th>
<th>$z_0$ (mm)</th>
<th>$z_d$ (mm)</th>
<th>$P(\text{drift})$</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.09</td>
<td>0.000</td>
<td>0.0</td>
<td>-</td>
<td>0.95</td>
</tr>
<tr>
<td>2</td>
<td>0.19</td>
<td>0.03</td>
<td>0.0</td>
<td>-</td>
<td>0.94</td>
</tr>
<tr>
<td>3</td>
<td>-</td>
<td>0.0</td>
<td>0.0</td>
<td>0.03</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>-</td>
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<td>0.0</td>
<td>0.02</td>
<td>-</td>
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<td>0.55</td>
<td>0.58</td>
<td>0.0</td>
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</tr>
<tr>
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<td>0.30</td>
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<td>0.21</td>
<td>0.98</td>
</tr>
<tr>
<td>7</td>
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<td>8</td>
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</tr>
<tr>
<td>9</td>
<td>0.86</td>
<td>0.76</td>
<td>0.0</td>
<td>0.92</td>
<td>&gt;0.99</td>
</tr>
</tbody>
</table>

These velocity profiles are notably noisier than those shown in Figures 3.7 and 3.9. The first two profiles are all but unusable, showing very low $z_0$ and very high distortion. There are several reasons for the noise. One is that the pressure transducer was exposed to variations in temperature which may cause small fluctuations in measured values. Another is the relatively long length of tubing used to connect the transducer to the rake, compared to measurements over the wooden floor. This might only cause small errors, but these are more noticeable at small velocities and hence low dynamic pressures.

At higher wind speeds, a more pronounced influence may be the significant quantities of snow drifting in the tunnel; Table 3.4 also gives the mean frequency of drift at 50 mm above the surface that was observed in each plateau. Drifting snow impacts on the pressure taps, causing fluctuations, and alters the boundary layer near the surface. The drift is not constant, but intermittent, and so the modification to the boundary layer also changes, and the height that drift reaches also changes. Also, because the measurements over the height of the rake are not instantaneous or simultaneous, but sequential, there
is a good possibility that this is reflected in transients near the surface. However, above the drifting material, the boundary layer should be influenced only by the modification of the apparent roughness, as was noted by Owen (1964).

These results suggest that the MiniAirs at 50 and 100 mm above the surface should be in the logarithmic region, and that measurements around the drift threshold should be unaffected by the low frequency of drift.

3.5 Uncertainty in $u_*$ and $z_0$

Uncertainty is usually described as the standard deviation of repeated results around a mean value. This can be estimated for a single measurement if the standard deviation of the input data is known. In this case, $u_*$ and $z_0$ are calculated from the mean velocity at two heights in the boundary layer, using the linear fit described in Section 2.1. Assessing the uncertainty in those parameters therefore requires the uncertainty in velocity, $\delta u(z)$ and height, $\delta z$ to be known. This is assuming that the Von Karman constant $\kappa$ is known.

From Equation (2.1), it can be seen that the uncertainty in $u_*$ is a function of the measurement height and the velocity at that height, and is not constant. The uncertainty must then be calculated for each separate measurement.

To estimate the uncertainty in a single measurement, the 'jitter' procedure of Moffat (1982) was used;

1. calculate ideal velocities at the measurement heights ($i = 1, 2, ..., n$) using a chosen $u_*$ and $z_0$.

2. calculate the uncertainty at each height in the profile by perturbing each point in the profile;

- calculate the error due to the height $z_i$ by calculating $u_*$ and $z_0$ from a fit to the displaced profile;

$$E(u_*, z_i) = \left( \frac{u_*(z(i) + \delta z) - u_*(z(i) - \delta z)}{2} \right)^2$$

and

$$E(z_0, z_i) = \left( \frac{z_0(z(i) + \delta z) - z_0(z(i) - \delta z)}{2} \right)^2$$

- calculate the error due to the velocity at height $z_i$ by calculating $u_*$ and $z_0$ from a fit to the accelerated or decelerated profile;

$$E(u_*, u(z_i)) = \left( \frac{u_*(u(z_i) + \delta u) - u_*(u(z_i) - \delta u)}{2} \right)^2$$

and

$$E(z_0, u(z_i)) = \left( \frac{z_0(u(z_i) + \delta u) - z_0(u(z_i) - \delta u)}{2} \right)^2$$

where $\delta u$ is the uncertainty from the velocity measurement.

3. calculate the absolute uncertainty for the complete profile from $\delta u_* = \sqrt{\sum_{i=0}^n E(u_*)}$ and $\delta z_0 = \sqrt{\sum_{i=0}^n E(z_0)}$.

4. calculate the relative uncertainty from $\delta u_*/u_*$ and $\delta z_0/z_0$. 


### 3.5.1 Measurements with MiniAirs

Profiles were initially measured using two MiniAir anemometers at 0.05 and 0.1 m above the snow surface. The uncertainty in the velocity measurement in this case is the largest of 0.5% of the full scale (20 m s\(^{-1}\)) or 1.5% of the expected reading at \(z_i\). This assumes that there is zero bias in the readings, no overall change in the wind tunnel behaviour during a measurement interval and that the data acquisition system resolution does not truncate the readings, and so the random error of the measurements is that given by the manufacturer. Velocity fluctuation data (as shown in Figure 3.4) cannot be used to estimate this value, as that data includes fluctuations induced by real physical processes, in that case turbulence. A cross-calibration of the MiniAirs at the same location in the tunnel showed that the measured velocities agreed to ±1% with a correlation coefficient of >0.99 at velocities from 3-15 m s\(^{-1}\).

Measurement heights (\(z_0\)) were measured before the start of an experiment using a ruler; the heights are thus known to a precision of ±1 mm, at least until erosion occurs.

The results for a two-point profile using the MiniAirs are shown in Figure 3.11, in the range \(0.1 \leq u_* \leq 1 \) m s\(^{-1}\) and \(1 \times 10^{-6} \leq z_0 \leq 1 \times 10^{-2} \) m.

**Figure 3.11:** Contours of relative uncertainty for 2-point MiniAir profiles. (a) \(\delta u_*/u_*\) and (b) \(\delta z_0/z_0\), for the likely range of \(u_*\) and \(z_0\) in wind tunnel experiments, assuming anemometers at 0.05 and 0.1 m above the surface.

Uncertainty in \(z_0\) of more than 100% at \(z_0 = 1 \times 10^{-4}\) m is relatively large. However, as was shown in Table 2.1, \(z_0\) changes between surfaces by several orders of magnitude, and so the accuracy of the measurements is sufficient to at least determine the relative magnitude of \(z_0\) for snow compared to other surfaces. It should then be possible to assess the value of \(C\) in equation (2.20) to ≈ 100%.

The use of two probes at 5 and 10 cm is also a compromise between needing the velocity near the particle counter, and having to stay within the limits of the logarithmic region and the boundary layer which has developed over the snow. Increasing the accuracy of the anemometers would decrease the uncertainty in \(u_*\) and \(z_0\), but during the development of the wind tunnel we found no sensors which offered the same accuracy, robustness and ease of use.

### 3.5.2 Measurements with a dynamic pressure rake

The velocity profile in the boundary layer was also measured using a dynamic pressure rake. The rake extended from approximately 3 mm above the surface to 0.25 m, but only the bottom 10 cm of the boundary layer was found to be in the logarithmic region. Uncertainty in the velocity measurements in this case was due to uncertainty in the dynamic pressure, which was measured as a differential pressure between the rake and the static port of a pitot-static tube. The differential pressure was measured using an MKS Baratron 120 pressure sensor (see Section 3.4). The uncertainty of the measured pressure was...
assumed to be the manufacturer’s quoted uncertainty, 0.12% of reading, and this was used to generate an error in velocity from Bernoulli’s Equation. Losses due to leaks were neglected as all tubes were assumed sealed. Pressure tap heights were known to ±1 mm.

The results for an 8-point profile using dynamic pressure measurements are shown in Figure 3.12 in the range $0.1 \leq u_\ast \leq 1$ m s$^{-1}$ and $1 \times 10^{-6} \leq z_0 \leq 1 \times 10^{-2}$ m.

![Figure 3.12: Contours of relative uncertainty using rake profiles. (a) $\delta u_\ast/u_\ast$ and (b) $\delta z_0/z_0$, for the likely range of $u_\ast$ and $z_0$ in wind tunnel experiments. Analysis is based on measurements of dynamic pressure at 6, 16, 26, 46, 56, 66, 86 and 116 mm above the surface.](image)

Other errors are also possible in $z$ and $u(z)$ as a result of turbulence or flow displacement around the pitot tubes. These errors were calculated to be smaller than the uncertainty in $z$ and $u(z)$ than was assumed for the error propagation, and so are not considered separately.

Uncertainty in $u_\ast$ is approximately constant at about 4%, while uncertainty in $z_0$ is almost an order of magnitude worse. However, the rake is not ideally suited to measurements over snow because of the measurement method. The dynamic pressure at each tap is measured sequentially by connecting the tap to the positive port of the MKS Baratron 120D pressure sensor via a Scanivalve. The static pressure is always connected on the negative side of the pressure sensor. The measurement at each port takes some 20 seconds, plus a switching time and some time for the pressure in the tube between the Scanivalve and pressure sensor to equalise to the measured pressure. In total, an 8-height measurement takes about 4 minutes, which increases the length of an experiment and leads to the risk that the snow surface will be modified by the sustained wind. The rake is the preferred measurement method when time is not important, for example during measurements over analogous materials.

The rake is also used to establish the relationship between $u_\ast$ and $z_0$ after drift occurs, after Owen (1964). The accuracy of $C$ in Equation (2.20), where $z_0 = C u_\ast^2 g$, is therefore also a function of the instrumentation and setup. Applying the same routines as used to calculate the uncertainty in $u_\ast$ and $z_0$ gives the result shown in Figure 3.13. This shows that the system is sufficient to resolve $C$ to approximately ±20% at the expected $z_0$ for snow.

### 3.5.3 Profile comparisons

The rake and MiniAir systems are both within 25 cm of the centre-line of the tunnel, and the rake is about 40 cm upstream of the MiniAirs. The two systems are offset by 10 cm to minimise rake wake effects on the MiniAirs. Ideally the rake and MiniAir systems should give the same velocities and hence $u_\ast$ and $z_0$ during an experiment. With no drift, or only low drift rates ($P(drift) \leq 0.1$), $u_\ast$ varied by less than 20% between the rake and MiniAirs, and $z_0$ by less than 150%; this is what would be
expected from the error bands associated with the two measurement systems. In almost all cases, the rake showed higher $u_*$ and $z_0$ than the MiniAirs. This resulted from acceleration at the lowest point in the MiniAir profile, caused by a slight pressure gradient downstream of the rake. As the rate of drift increases, the differences in velocities measured by the two systems becomes larger, increasing to about 50% difference in $u_*$ and in some cases, several orders of magnitude difference in $z_0$. As drift increases, it will influence the velocity measured at the height of the lowest anemometer, but this influence is difficult to assess or detect. Measurements with the rake show the influence more clearly, and so it may be expected that the rake measurements are more reliable when sustained drift occurs.

### 3.5.4 Repeatability

Measurements from both the pressure rake and the MiniAirs show good repeatability. MiniAir velocities varied by less than 1%, when the mean velocities are compared between the first $n/2$ and last $n/2$ measurements of a plateau of $n$ measurements. In the absence of being able to repeat an experiment over a specific surface, this suggests that velocity measurements made within a short time of each other are repeatable, and vary within the range of accuracy of the MiniAirs themselves. Rake profiles included repeated measurements at one point in the profile, which were usually within 1% of a mean, and often better than 0.2%. Together, this suggests that the device uncertainty is a good proxy for the total uncertainty in a measurement.

A series of 3 experiments were made in March 2006 using sieved snow which was prepared to a similar standard on each occasion. The prepared snow samples had the same depth, were made using the same sieves and preparation methods, and ambient temperatures on each occasion were similar, in the range $-3 < T_a(°C) < 0$. Experiments used the same ventilator speeds, and approximately the same length of time at each measurement plateau. The variation in the free-stream velocity during a measurement plateau was less than 1%. Results are shown in Figure 3.14 $u_*$ and $z_0$ were calculated from velocities measured by the dynamic pressure rake. Fits were limited to data where $z_0 < 0.1$ m. Fits to data with drift were limited to $0.05 < z < 0.1$ m.

Although the tests over snow were supposed to be similar, results from the snow characterisation and measurements from the instruments within the wind tunnel show that there may be some variation in both the snow and conditions between the tests. These are summarised in Table 3.5.

Grain diameter, snow density, mapped sphericity and dendricity all change from test to test. As these parameters influence the shape of the snow particles and the degree of flow through the snow, the large variation between tests makes it impossible to say that surface conditions were identical, and variations in $z_0$ between snow surfaces are larger than the variation found in repeated measurements.
3.5. UNCERTAINTY IN $U_*$ AND $Z_0$

Figure 3.14: Repeated measurements of $u_*$ and $z_0$ over sieved snow. Data is classified as no drift (+), or drift observed (○).

Table 3.5: Weather and snow characteristics during sieved snow tests in March 2006.

<table>
<thead>
<tr>
<th>Date</th>
<th>Ambient $T_a$ ($^\circ$C)</th>
<th>RH (%)</th>
<th>Density (kg m$^{-3}$)</th>
<th>$d_l$ (mm)</th>
<th>$d_h$ (mm)</th>
<th>SP</th>
<th>DN</th>
</tr>
</thead>
<tbody>
<tr>
<td>21/03/2006</td>
<td>-0.4</td>
<td>68</td>
<td>424</td>
<td>1.15</td>
<td>1.06</td>
<td>0.97</td>
<td>0.00</td>
</tr>
<tr>
<td>22/03/2006</td>
<td>-0.8</td>
<td>73</td>
<td>401</td>
<td>1.16</td>
<td>1.21</td>
<td>0.67</td>
<td>0.00</td>
</tr>
<tr>
<td>23/03/2006</td>
<td>-0.5</td>
<td>72</td>
<td>522</td>
<td>1.15</td>
<td>1.05</td>
<td>0.51</td>
<td>0.00</td>
</tr>
</tbody>
</table>

over foams. There is also a trend towards increasing $z_0$ as drift starts.

3.5.5 ‘Patchy’ floor

A potential drawback of the wind tunnel arrangement is that the surface is by necessity ‘patchy’. That is, sections of differing surface roughness (roughness elements, carpet and then snow) follow each other. A boundary layer which develops over a surface of a specific roughness length and then encounters a second surface roughness, will adapt to the new roughness. This adaption takes the form of an internal boundary layer growing from the upstream edge of the second patch. This is shown schematically in Figure 3.15 where a second surface with a lower $z_0$ is encountered, causing the velocity measured within the new boundary layer to increase, compared to the upstream surface. If one anemometer is still outside of the internal boundary layer, the apparent $z_0$ as measured from the $z$-intercept will be lower than the real value for the downstream surface.

The interface height of the internal boundary layer, $z_i$, is a function of the roughness length of the second surface, in this case snow, and the fetch over which the new boundary layer develops, $x$. Arya (2001, pp. 326-327) reports that experimental data shows

$$z_i = z_0 a_i \left( \frac{x}{z_0} \right)^{0.8}$$

where $a_i$ is an empirical constant, $0.35 \leq a_i \leq 0.75$. The highest anemometer was mounted 0.1 m above the surface after fetch $x = 3$ m; in the worst case $a_i = 0.35$ and the smallest roughness length at
Figure 3.15: Schematic of the boundary layer over a patchy surface. In this case, the upstream surface has a larger $z_0$ than the downstream surface. The transition to the new surface causes an acceleration near the surface, and so $z_0$ measurements in the transition region are lower than the actual values, measured further downstream.

which $z_i \geq 0.1$ m is $z_0 = 2.4 \times 10^{-5}$ m. Roughness lengths smaller than this were therefore rejected as being contradictory, as they would require the upper anemometer to be above the interface. Figure 3.8 suggests that this approach correctly identified the limit to the internal boundary layer, giving a good agreement between solid wall boundary layers and those given in literature.

3.6 Drifting snow measurements

Wind tunnel motor rotation speeds are increased between distinct plateaux. This increases the free-stream velocity accordingly, and thus $u_*$. During the periods of acceleration between plateaux, the saltation system responds to the acceleration and resulting increase in shear, causing rapid increase in drift activity. This drift activity was measured by the SPC at a nominal height of 50 mm above the snow surface. A typical time series of the rate of drift and the corresponding 50 and 100 mm velocity is shown in Figure 3.16.

Literature suggests the time delay between an acceleration in the flow and drift activity should be in the range 1-5 seconds (Anderson and Haff, 1988). Testing this hypothesis would have been possible by calculating the cross-correlation between two time series with an introduced lag time. However, this would have required a time resolution at least twice as high as the expected time delay, i.e. 2 Hz. This sampling frequency was not possible with the SPC or MiniAir anemometers. Because of the 1 Hz or slower sampling interval of the instrumentation in the wind tunnel, the results presented here are focused on longer time-periods and use mean values of the velocity profiles, SPC mass fluxes and particle counts.

Wind profiles were measured using the MiniAirs only while they were unaffected by drift. Drift causes significant surface erosion, increasing error in $z_0$ and moving the SPC relative to the surface. Drifting snow also influences the velocity measured by the MiniAirs, changing the measured $u_*$. In experiments where more than 5 mm erosion was observed, only the first 3 plateaus with drift were used. Because wind speeds are increased stepwise, it is assumed that the initial drift which is measured is solely due to aerodynamic forces acting on the snow surface.

An indication of the stability of the system is the drift activity during the measurement plateaux. As was noted in Section 2.3 increased boundary layer shear will result in increased particle flux. If the wind tunnel velocity changes significantly during a measurement plateau, a change in saltation would be expected, followed by a gradual relaxation to a new equilibrium level. Changes in drift activity might also be expected if snow layers of considerably different properties were exposed by erosion of the surface. There is some suggestion of transients in activity during the plateau in the data shown in Figure 3.17 at low $u_*$, but as $u_*$ increases the drift activity becomes more homogeneous.
3.6. DRIFTING SNOW MEASUREMENTS

Figure 3.16: Raw wind speed and drift data during a test. Data is taken from the test on 9th March 2005. Every 10th velocity measurement is shown, all drift data is shown.

Figure 3.17: Wind speeds and drift activity during a test. Data is taken from the test on 9th March 2005. Velocities were measured using MiniAirs. The number of drifting particles per second was measured by the SPC. Time series are normalised by the duration of the measurement plateau ($\Delta T$). The data used to calculate $u^*$ and $z_0$ are limited to $0 \leq t/\Delta T \leq 1$, and values of $u^*$ and $P(drift)$ are given for the plateau.

An indication that the windowing of the data into the plateau does not cause a distortion of the data comes from the particle sizes of drifting snow. The size distributions measured by the SPC are shown in Figure 3.18. The upper plot shows the particle size distribution that was measured in the interval from the start of the experiment, to the end of the fifth plateau. Sizes measured in the transients between plateaux are also included. The lower plot shows the size data that was recorded when the wind speeds were constant, and transients are excluded. The transients included about 15% of the total
number of particles that were in motion. The result of the windowing to remove the transients is to remove a few of the larger particles that were in motion, but otherwise the size distribution appears to remain constant.

![Graph](image)

**Figure 3.18:** Drifting particle size distribution before and after plateau selection. Top; raw drifting particle size distribution as measured by the SPC. Bottom; particle size distribution after plateau selection. Both data sets are from the test on 9th March 2005. The line is a fit to the distribution using a gamma fit (Equation 3.3).

The probability distribution function of grain sizes is usually defined by the gamma distribution (Budd et al., 1966; Schmidt, 1982; Nishimura and Nemoto, 2005). Several forms of this have been given in literature; the following is used in this document. The probability of observing a grain of diameter \(d\) is given by

\[
P(d) = \frac{d^{\alpha-1}}{\beta^\alpha \Gamma(\alpha)} e^{d/\beta}. \tag{3.3}
\]

The mean grain size is given by the product \(\alpha\beta\). The variance of the grain diameter is given by \(\alpha\beta^2\).

The effect of windowing the data on the parameters of the gamma size distribution is summarised in Table 3.6. The influence is minimal, resulting in no practical change in size characteristics. The biggest influence on the particle sizes is the mean \(u_*\); further analysis (not shown here) showed that if the measurements at higher \(u_*\) were included despite their increased uncertainty, the mean particle diameter would increase. This follows from the capability of the flow to transport larger particles at higher \(u_*\) (Bagnold, 1941; Owen, 1964).

**Table 3.6:** Drifting particle size distribution before and after plateau selection. The gamma probability distribution parameters \(\alpha\) and \(\beta\) are given for data plotted in Figure 3.18. The mean particle diameter is \(d = \alpha\beta\).

<table>
<thead>
<tr>
<th>Data</th>
<th>(\alpha)</th>
<th>(\beta)</th>
<th>(\overline{d} (\mu m))</th>
</tr>
</thead>
<tbody>
<tr>
<td>all</td>
<td>4.53</td>
<td>29.95</td>
<td>135.6</td>
</tr>
<tr>
<td>windowed</td>
<td>4.43</td>
<td>30.76</td>
<td>136.3</td>
</tr>
</tbody>
</table>
3.7 Summary

Instrumentation has been set up in a wind tunnel to allow simultaneous measurements of wind profiles and drift activity over a natural snow surface. Time-averaged wind profiles measured using anemometers and a pressure rake show that a logarithmic boundary layer exists over the snow. Once drift starts, a logarithmic profile still exists above the saltating region, while below it there is a departure from the logarithmic profile. The accuracy of the measurements is sufficient to observe the influence of the surface on $z_0$ and obtain an estimation of the ratio $C = 2g z_0/u_*^2$ to within 25%, which is better than field investigations have been able to achieve. Repeated measurements over sieved snow show differences in $z_0$ which may be a result of snow properties. The time response of the instrumentation is insufficient to investigate the time response of drift, but the drift activity appears to be in equilibrium when wind speed is constant.
Chapter 4

The drift threshold

I used to be Snow White, but I drifted.

Mae West (1893-1980)

In this chapter a definition for the drift threshold is given, and methods are proposed for estimating it from measurements. A method based on the overall nature of drift is selected as the most useful for both field and laboratory work, and the data is analysed accordingly. The impact of ambient conditions on the threshold is considered. Several other quantitative indicators are also investigated. Finally, the alteration of the boundary layer by drift is shown. Much of the data presented in this chapter appeared in Clifton et al. (2006). The drift mass flux data has since changed slightly following a recalibration of the SPC (see section A.1), but this has not resulted in any significant difference from the conclusions presented in Clifton et al. (2006).

4.1 Defining a threshold

The drift threshold could be assessed by eye from the drift activity in the tunnel, or by ‘gut instinct’ from plots of mass flux as a function of reference wind speed. However, both methods are subjective. In the following section, a method is described to establish \( u^*_{\text{t}} \), which reflects the physics of the boundary layer flow, the drift threshold and the processes which occur at lower and higher \( u^* \). This requires time-averaged data and regression using physical models.

The threshold velocity is defined for a flat, smooth and uniform snow surface. It is defined using measurements from a snow particle detector mounted 50 mm above the surface. Measurements at 50 mm are common within aeolian literature (Schmidt, 1982; Gillette et al., 1997; Nishimura and Nemoto, 2005) and represent a height that saltating particles can reach, but will not be attained by reptating particles. If there is no precipitation, particles only reach heights above 100 mm when there is significant suspension, which requires a higher \( u^* \) than saltation (Nishimura and Hunt, 2000). Thus 50 mm is a compromise height which is insensitive to reptation, but sensitive to saltation before suspension occurs.

The friction velocity \( u^* \) is used as the reference velocity, as this is independent of the measurement height, giving a threshold value \( u^*_{\text{t}} \). \( u^* \) is measured above the saltating material, and is a time-averaged value over an interval \( O (100 \text{ s}) \), much longer than the flight time of an individual saltating particle \( O (0.1 \text{ s}) \). A regression technique using all data from a test is preferred, as this forces the threshold condition to be representative of the whole test, not just a single event such as the \( u^* \) of the plateau at which drift first occurs.

4.1.1 Mass flux of drifting particles

The total horizontal mass flux \( Q \) from a loose surface has been analysed and measured for sand and snow (by Bagnold, 1941, pp. 69-70, Shao and Raupach, 1992 and Nishimura and Nemoto, 2005 amongst others) and results give a relationship of the form

\[
Q = \begin{cases} 
0 & \text{for } u^* < u^*_{\text{t}} \\
 a_m (u^* - u^*_{\text{t}})^3 & \text{for } u^* \geq u^*_{\text{t}} 
\end{cases}
\]

(4.1)
From dimensional considerations, the constant of proportionality $a_m$ is given by $a_m = \frac{\lambda \rho d}{g}$, where $\alpha_m$ is some constant for a particular surface.

Measurements with the SPC in the SLF wind tunnel do not cover the entire saltation layer. The SPC measures the flux over a 2.5 mm high window, positioned 50 mm above the snow surface. The effect of moving the saltating cloud past the SPC has to be considered. Separating the horizontal mass flux at a height $z$ into a total mass flux $Q(u_*)$ and a profile function $\Phi(u_*, z)$ which defines the form of the drifting flux, we can define a general form of the mass flux at a height, $q(z)$, as a function of $u_*$ and $z$, which for $u_* > u_{st}$ is given by

$$q(z) = Q(u_*)\Phi(u_*, z). \quad (4.2)$$

Various experiments have shown that the mass flux of drifting snow decays exponentially with height (Budd et al., 1969; Maeno et al., 1995; Nishimura et al., 1998; Nishimura and Hunt, 2000). The saltating flux at the surface must be zero, and hence there is a height at which the mass flux is at a maximum. This is not seen in field measurements, because of the difficulty of achieving the necessary resolution close to the surface and the effect of suspension. Pomeroy and Gray (1990) give $h_s = 1.6u_*/g$, which would be 12 mm at $u_*=0.4 \text{ m s}^{-1}$, and so the SPC is above the height of maximum flux. The total flux for $z < h_s$ is assumed to be negligible compared to the mass flux for $z > h_s$.

If the measurement height is $z$, and the characteristic height of the saltation system is $L$, and the integral is 1, so that the profile function scales the total mass flux, the profile function is given by

$$\Phi(u_*, z) = \frac{1}{L} e^{-z/L}. \quad (4.3)$$

The characteristic height $L$ can be estimated from conservation of energy. If a particle leaves a surface with a vertical velocity $\eta u_*$, where $\eta$ is some constant, conservation of kinetic and potential energy dictates that it will reach a height given by $\frac{1}{2g}(\eta u_*)^2$, if external forces are zero. The parameter $2/\eta^2$ is defined as $\lambda$, where $\lambda$ is a dimensionless parameter for a single experiment. Hence, $L$ is given by

$$L = \frac{1}{\lambda g} u_*^2. \quad (4.4)$$

Nishimura and Hunt (2000) report $\lambda = 0.45$ for drifting 0.48 mm diameter snow, and 0.13-0.16 for drifting ice particles.

Substituting Equation (4.4) and (4.3) into Equation (4.2) gives the mass flux at a single height when $u_* > u_{st}$;

$$q(z) = a_m (u_* - u_{st})^3 \frac{\lambda g}{u_*^2} e^{-\lambda z g / u_*^2}. \quad (4.5)$$

### 4.1.2 The threshold model

Regression is used to identify the threshold from mass flux and friction velocity data. By combining equations (4.5) and (4.1), $u_{st}$ can be found by fitting the mass flux - friction velocity data to the following equation, using a least-squares error minimisation routine.

$$q(z) = a_m (\max(0, u_* - u_{st}))^3 \frac{\lambda g}{u_*^2} e^{-\lambda z g / u_*^2}. \quad (4.6)$$

The unknowns in Equation (4.6) are $a_m$ and $u_{st}$. All other data is measured during the experiment. A fit to data from a test is shown in Figure 4.1 using 3 different values of $\lambda$.

Figure 4.1 shows that the influence of $\lambda$ is small, and does not influence $u_{st}$, and so it is assumed that $\lambda = 0.45$ can be taken for all data, as was found by Nishimura and Hunt (2000) for snow. The regression technique is also tolerant of inaccuracy in the actual value of $q(z)$; a sensitivity test to simulate possible SPC measurement errors was carried out by increasing and decreasing the mass flux by a factor of 10. This resulted in no change in the value of $u_{st}$.

Figure 4.1 also demonstrates the reason for this method; if the threshold had been defined as the first point where drift was detected, the threshold would have been at $u_* < 0.32 \text{ m s}^{-1}$, and would have been defined by a single data point, which from Figure 4.16 may be a transient event. Less particles were
4.1. DEFINING A THRESHOLD

Figure 4.1: Drifting snow mass flux as a function of $u_*$. Snow mass flux is averaged over measurement plateau during measurements on March 9th 2005. Data is fitted using Equation (4.4). 3 different values of $\lambda$ were tested, and $u_* = 0.30$ m s$^{-1}$ in each case. The numbers beside each data point are the total number of observed particles at each point.

detected during the next measurement plateau than before, despite the increase in $u_*$, which would normally result in an increase in the amount of drift. Looking at the whole data series does suggest a change at $u_* \approx 0.3$ m s$^{-1}$, despite the drop in drift after the first observation. The $u_* t$ from the regression was 0.3 m s$^{-1}$, in this case. Using a regression to all data in a test, the threshold is then defined in terms relevant to mass movement, not by a single measurement or small number of drifting particles.

4.1.3 The time-fraction equivalence method

By definition, the probability of drift occurring at $u_* < u_* t$ is zero. Above the threshold, the average chance of observing a particle in motion during a long interval is non-zero, so long as the interval is much longer than the particle flight time. Until $u_*$ is high enough to move the most common particles in a snow pack to the height of observation, drift is likely to be intermittent as the supply of suitably sized particles is limited. There is also the potential that drift might cease completely, if all particles of small enough size to drift, are eroded. It could be argued that this case represents a new surface, and short measuring times were used in these experiments to avoid that situation.

The instantaneous momentum transport by wind during an interval is not constant, and could fall below the drift threshold value. The percentage of time that the wind speed exceeds the threshold can be described as a wind intermittency, $P(u)$. Similarly, the drift may also be intermittent, depending on $u(z)$, and may occur only for a certain percentage of time. This gives the drift intermittency, $P(\text{drift})$. The Time Fraction Equivalence Method (TFEM, see Stout and Zobeck, 1996; Wiggs et al., 2004; Schönfeldt, 2004) uses these intermittencies to calculate the drift threshold wind speed. The threshold is defined as that wind speed which minimises the difference between the wind and drift intermittencies for an observation period. TFEM thresholds are usually defined with reference to a wind speed measured at a given height, rather than $u_*$, but in principle either value could be used.

Figure 4.2 shows the probability of drift during the plateau in an experiment, where the probability of drift during each plateau, $P(\text{drift})$, is defined as the total number of observations of drift, in seconds, divided by the duration of the plateau. The Time Fraction Equivalence Method (TFEM) uses these intermittencies to calculate the drift threshold wind speed. The threshold is defined as that wind speed which minimises the difference between the wind and drift intermittencies for an observation period. TFEM thresholds are usually defined with reference to a wind speed measured at a given height, rather than $u_*$, but in principle either value could be used.
Chapter 4. THE DRIFT THRESHOLD

of the TFEM. The residuals increase as the probability tends to 0 or 1, suggesting that a higher order fit is needed. Therefore, it is assumed that the drift activity is driven by excess shear, i.e., that amount of wind shear above the threshold. This would give $P(\text{drift}) \propto (u^*_t)^2 - u^*$. A fit using this function is also shown in Figure 4.2. In this case, the use of a more complex model did not improve the quality of the fit; the coefficient of determination was $R^2 = 0.985$ for both fits.

The TFEM might be ideal for field measurements as it requires only simple measurement devices which register the presence (or not) of drift in an interval, and a wind sensor. However, as Figure 4.2 demonstrates, regression with small data sets is difficult, and in this case (and others, not shown here) leads to an over-prediction of $u^*_t$ by some 30%. The TFEM also ignores the extra data that can be gained by using an instrument like the SPC, which also delivers the mass flux of drifting material.

4.2 Threshold measurements

The data described here were obtained over two winters of measurements. A total of 22 experiments gave 15 clear thresholds which are presented here. Experiment conditions are described in Table 4.1 and the associated drift threshold data are summarised in Table 4.2. The drift thresholds were obtained by regression of the mass flux and $u^*_t$ data in the different plateaux during an experiment, using equation (4.6). Because $u^*_t$ is obtained from a fit to Equation (4.6), it does not correspond directly to a measurement plateau, and so has no $z_0$ measurement associated with it. Therefore, $z_0$ at $u^*_t$ is taken as the linearly interpolated value from the $u^*_t$ and $z_0$ of the measurements immediately around the threshold. The data is bracketed in the range $[u^*_t]$ and $[z_0]$ using the friction velocities and roughness lengths measured during the neighbouring measurements. Ambient conditions at the onset of drift are taken as the mean values in these plateau.

Snow from the surface of the trays was characterised before the trays were placed in the tunnel. Results are summarised in Table 4.1. The snow density, temperature and grain types were measured using a typical ‘field’ method after Colbeck et al. (1990). Photomicrographs were also taken and the grain forms determined using image processing. The image processing gave the particle sphericity $SP$ and dendricity $DN$ (Lesaffre et al., 1998). The process of snow metamorphosis results in a change of

Figure 4.2: Probability of drift occurrence during a test. Data is taken from the test on 9th March 2005. Fits to data using an assumption of linear dependence on $u^*_t$ ($u^*_t = 0.38 \text{ m s}^{-1}$) and shear-driven drift activity ($u^*_t = 0.38 \text{ m s}^{-1}$) are shown. $u^*_t$ is the point at which the fit departs from zero drift probability.
4.2. THRESHOLD MEASUREMENTS

$DN$ from 1 for new snow, towards zero for older snow. It is possible to distinguish between the two major processes of metamorphosis by describing snow with $DN < 0.5$ and $SP > 0.5$ as snow which is becoming rounded, and snow with $DN < 0.5$ and $SP < 0.5$ as snow which is becoming faceted. The process of metamorphosis, and the dependency of new snow on atmospheric conditions, is shown in Figure 4.5.

Table 4.1: Weather conditions and snow characterisation for wind tunnel tests.

<table>
<thead>
<tr>
<th>No.</th>
<th>Date</th>
<th>$T_a$ (°C)</th>
<th>RH (%)</th>
<th>Types*</th>
<th>Density (kg m$^{-3}$)</th>
<th>SMP Force (N)</th>
<th>$d_f$ (mm)</th>
<th>SP</th>
<th>DN</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>17/01/2005</td>
<td>-3.5</td>
<td>58</td>
<td>5a†</td>
<td>301</td>
<td>-</td>
<td>2.17</td>
<td>0.00</td>
<td>0.02</td>
</tr>
<tr>
<td>2</td>
<td>19/01/2005</td>
<td>-5.0</td>
<td>71</td>
<td>1cr, 1dr</td>
<td>98</td>
<td>-</td>
<td>1.02</td>
<td>0.36</td>
<td>0.63</td>
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<tr>
<td>3</td>
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<td>97</td>
<td>1f</td>
<td>162</td>
<td>0.04</td>
<td>0.58</td>
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<td>0.15</td>
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<tr>
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<td>1a, 1d</td>
<td>91</td>
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<td>0.37</td>
<td>0.27</td>
<td>0.13</td>
</tr>
<tr>
<td>5</td>
<td>25/01/2005</td>
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<td>1a, 1d</td>
<td>91</td>
<td>0.01</td>
<td>0.37</td>
<td>0.27</td>
<td>0.13</td>
</tr>
<tr>
<td>6</td>
<td>27/01/2005</td>
<td>-15.9</td>
<td>85</td>
<td>1e, 1a</td>
<td>99</td>
<td>0.01</td>
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<td>0.64</td>
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</tr>
<tr>
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<td>1d, 1c</td>
<td>88</td>
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<td>0.20</td>
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<tr>
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<td>1d, 1c</td>
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<td>2.02</td>
<td>0.26</td>
<td>1.00</td>
</tr>
<tr>
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<td>14/02/2005</td>
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<td>1b, 1d</td>
<td>83</td>
<td>0.03</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>10</td>
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<td>-7.9</td>
<td>76</td>
<td>2b</td>
<td>111</td>
<td>0.01</td>
<td>0.46</td>
<td>0.60</td>
<td>0.19</td>
</tr>
<tr>
<td>11</td>
<td>08/03/2005</td>
<td>-5.1</td>
<td>68</td>
<td>1c, 1d</td>
<td>76</td>
<td>0.01</td>
<td>1.15</td>
<td>0.22</td>
<td>0.88</td>
</tr>
<tr>
<td>12</td>
<td>09/03/2005</td>
<td>-4.0</td>
<td>66</td>
<td>1cr, 1f</td>
<td>31</td>
<td>0.01</td>
<td>0.96</td>
<td>0.47</td>
<td>0.57</td>
</tr>
<tr>
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<td>53</td>
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<td>0.48</td>
<td>0.40</td>
</tr>
<tr>
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<td>28/02/2006</td>
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<td>51</td>
<td>1d, 2a</td>
<td>63</td>
<td>0.00</td>
<td>1.20</td>
<td>0.28</td>
<td>1.00</td>
</tr>
<tr>
<td>15</td>
<td>02/03/2006</td>
<td>-4.9</td>
<td>31</td>
<td>2a, 3a</td>
<td>74</td>
<td>0.00</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

*Nomenclature is taken from the International Snow Classification scheme of Colbeck et al. (1990). Suffix 'r' indicates significant riming.

Depth hoar was excavated from slopes around the wind tunnel.

Table 4.2: Boundary layer characteristics at the drift threshold

| No. | Date        | $z_{0f}$ (mm) | $|z_{0f}|$ (mm) | $u_{*f}$ (m s$^{-1}$) | $|u_{*f}|$ (m s$^{-1}$) | $u_f$ (10) (m s$^{-1}$) |
|-----|-------------|---------------|----------------|-----------------------|-----------------------|------------------------|
| 1   | 17/01/2005  | 0.12          | 0.11 - 0.12    | 0.69                  | 0.67 - 0.71           | 19.1                   |
| 2   | 19/01/2005  | 0.10          | 0.10 - 0.10    | 0.34                  | 0.30 - 0.37           | 9.4                    |
| 3   | 20/01/2005  | 0.09          | 0.06 - 0.11    | 0.38                  | 0.31 - 0.45           | 10.8                   |
| 4   | 25/01/2005  | 0.05          | 0.04 - 0.07    | 0.27                  | 0.25 - 0.29           | 8.0                    |
| 5   | 25/01/2005  | 0.12          | 0.09 - 0.15    | 0.34                  | 0.33 - 0.34           | 9.3                    |
| 6   | 27/01/2005  | instruments above $z_i$ |          |                      |                       |                        |
| 7   | 02/02/2005  | 0.07          | 0.05 - 0.08    | 0.34                  | 0.32 - 0.36           | 9.8                    |
| 8   | 03/02/2005  | instruments above $z_i$ |          |                      |                       |                        |
| 9   | 14/02/2005  | 0.08          | 0.07 - 0.08    | 0.28                  | 0.25 - 0.32           | 8.2                    |
| 10  | 24/02/2005  | 0.04          | 0.04 - 0.05    | 0.44                  | 0.40 - 0.48           | 13.3                   |
| 11  | 08/03/2005  | 0.13          | 0.12 - 0.14    | 0.38                  | 0.35 - 0.41           | 10.4                   |
| 12  | 09/03/2005  | 0.10          | 0.08 - 0.12    | 0.30                  | 0.25 - 0.34           | 8.4                    |
| 13  | 21/02/2006  | instruments above $z_i$ |          |                      |                       |                        |
| 14  | 28/02/2006  | instruments above $z_i$ |          |                      |                       |                        |
| 15  | 02/03/2006  | instruments above $z_i$ |          |                      |                       |                        |
4.2.1 Bulk snow properties

The influence of the bulk snow properties on the threshold \( u_* \) and \( z_0 \) is shown in Figure 4.3. Coefficients of determination from linear regressions between \( u_* \), \( z_0 \) and the snow properties are also given.

![Figure 4.3](image-url)

**Figure 4.3:** Influence of bulk snow properties on threshold conditions. Density is measured on site in the top 50 mm of snow, and the surface particle diameter is the mean from 30 or more images. Data are grouped by surface snow type; * new and decomposing snow \((DN > 0.5)\); ◆ rather faceted \((DN \leq 0.5, SP \leq 0.5)\); △ rather rounded \((DN \leq 0.5, SP > 0.5)\); + unclassified.

The highest observed correlation between snow properties and drift threshold parameters is between \( u_* \) and the snow density. In this case, \( R^2 = 0.82 \), but this could be spurious, given the strong influence of the high density and \( u_* \) of just one test (17th January 2005) on the resulting fit. The next highest correlation, \( R^2 = 0.16 \), was seen between \( u_* \) and \( d_i \), and may also be prone to influence from one test more than others. The low correlation (\( R^2 = 0.16 \)) between surface particle size and \( z_0 \) is unexpected, as other measurements over smooth granular surfaces show high correlations between \( z_0 \) and particle size (see Table 2.2). This suggests that occasionally observed surface topology may increase the roughness length, over and above that due to the granular structure of the snow.

4.2.2 Ambient conditions

Figure 4.4 also shows no strong correlation between the ambient conditions (temperature and humidity) and the roughness lengths or friction velocity at the threshold. Magono and Lee (1966) showed that cloud temperature and humidity change the form of fresh snow particles, which would then alter the surface structure (see Figure 4.5). The same figure also shows that the form of the snow crystals changes with age. Because of this, it might be expected that the temperature and humidity, or the directly observed crystal form, would influence \( z_0 \) of a flat snow cover. However, this is not seen in Figure 4.4. This is probably because the measured near-ground conditions do not correlate directly with cloud conditions, where the snow particles form. It may also be because the relationship between temperature, humidity and form is not linear, but follows a complex pattern (see Figure 4.5). Flakes which form at temperatures of \( T_a \leq -3 \) °C or \(-10 \leq T_a \leq -22 \) °C have similar plate-like forms, while those that form at \(-3 \leq T_a \leq -10 \) °C or less than -22 °C, tend to be columnar. There is no obvious trend towards two temperature dependent regimes of \( z_0 \) in Figure 4.4. There is also no clear distinction between \( z_0 \) for
4.2. THRESHOLD MEASUREMENTS

Figure 4.4: Influence of ambient conditions on threshold conditions. Data are grouped by surface snow type; * new and decomposing snow ($DN > 0.5$); ○ rather faceted ($DN \leq 0.5, SP \leq 0.5$); △ rather rounded ($DN \leq 0.5, SP > 0.5$); + unclassified.

new, faceted or rounded flakes identified from images.

A low correlation ($R^2 = 0.18$) was obtained between $u_\ast t$ and the ambient temperature. Kirchner et al. (2001) shows strain rates at which an ice matrix still behaves as a ductile material are higher at high temperatures compared to lower temperature. Assuming that strain rates at the snow surface are similar in all drift experiments, this would mean that a warmer surface would be less likely to fail and eject a particle, and so warmer surfaces would be expected to have higher $u_\ast t$. There is a trend toward this in Figure 4.4(a), but again the limited data makes it difficult to establish a relation.

Figure 4.5: Snow particle morphology and metamorphosis. (a) New snow particle morphology as functions of temperature and humidity. (b) Metamorphosis of snow described using sphericity $SP$ and dendricity $DN$.
4.2.3 Surface penetration resistance

Snow surface penetration resistance was measured before the experiments using a SnowMicroPen (SMP). Figure 4.6 shows the mean force applied by the SMP compared to the threshold shear, $\tau^{*} = \rho u^{*2}$. The mean force is the mean of readings when the SMP tip is between 4 and 5 mm into the snow, and represents complete immersion of the SMP tip. The SMP measures the total force required to break contact points and move grains within the snow. Aerodynamic entrainment at the surface during saltation is also a process where bonds are broken, and so a correlation between the aerodynamic and mechanical forces might be expected. However, as Figure 4.6 shows, the correlation is very low ($R^{2} < 0.01$).

![Figure 4.6: Surface penetration resistance compared to threshold aerodynamic shear stress. Penetration resistance was measured by SnowMicroPen.](image)

4.3 Comparison to threshold algorithms

Several algorithms have been developed for the threshold wind speed at which mass movement of granular material begins. This section presents some of the more commonly used approaches, and highlights issues with each of them.

4.3.1 Bagnold’s threshold parameter $A$

This is based on a force balance, as described in Section 2.2.1. Bagnold’s original data was from dry desert or beach sand. Sand which has been transported by wind or water, or weathered tends to be rounded or abraded, and may approximate spherical particles; see, for example, Figure 2.17 in Greeley and Iversen (1985). This also shows a Reynolds number dependency, with Bagnold observing that exceptionally smooth surfaces resisted erosion at $u_*$ high enough to erode particles much larger than those found on those surfaces.

Observed values of $A$, calculated using Equation (2.13) with particle diameters from image processing ($d_1$) are shown in Figure 4.8(a). This shows that the measured threshold $u_{st}$ are bounded by the predictions using the threshold parameter $0.1 \leq A \leq 0.2$, and do not appear to be Reynolds number dependent. The drift threshold for new and decomposing snow is grouped around $A \approx 0.1$, which is
4.3. COMPARISON TO THRESHOLD ALGORITHMS

a similar value to dry, granular material such as clover seeds, nut shells and sand (plotted in Figure 3.6 of Greeley and Iversen, 1985). By comparison, the drift threshold for older snow which is becoming rounded or faceted is higher than would be expected for similar sized sand in air, with $A \approx 0.17$. This may be a result of increased bond size in older particles compared to new snow, or may stem from different shapes which do not have the same geometry as new snow, which might then change the forces acting on a particle (see Figure 2.3). In general, snow particles are not spherical. This is clearly visible in the photomicrographs taken on March 9th 2005, for example, as shown in Figure 4.7.

![Figure 4.7: Outlines of surface snow forms from photomicrographs. Images were taken on 9th March 2005.](image)

A common engineering approach to correcting dimensions of non-spherical objects is to reduce the particle size to a hydraulic diameter, i.e. that diameter that an object would require, to have the same hydraulic resistance. The hydraulic diameter $d_h$ is defined as

$$d_h = \frac{4a}{p}, \quad (4.7)$$

where $a$ is the surface area of the object, and $p$ the perimeter. For a perfect sphere, this gives $d_h = d$. The measured value of the threshold parameter $A$ using $d_h$ is plotted in Figure 4.8(b).

The threshold parameter for snow using the hydraulic diameter is $A \approx 0.18$, and shows less scatter than seen using the mean diameter $d_I$. A possible reason for the increased threshold parameter of snow compared to dry sand is the inter-particle force. Several researchers have noted that the presence of water in sand and soil beds increases the threshold parameter to as high as 0.2, depending on the moisture content (e.g. Cornelis and Gabriels, 2003, 2004 and Ravi et al., 2006). This interstitial water is assumed to form a liquid bridge between particles, which increases the inter-particle force, raising $A$. In snow, a similar effect is seen from sintering of the ice particles, forming bonds between the grains.

While it appears that using the hydraulic diameter allows a better comparison of data from snow and other aeolian materials, the hydraulic diameter was not known a priori and had to be obtained from images. This approach is not ideal for a modelling situation, where only the grain geometric shape $d$ and the grain type might be known (see Figure 4.5). Can these then be used to determine the hydraulic diameter? From Figure 4.5, the dendricity $DN$ might be better expected to correlate with the increase of perimeter, as it represents the change from an open and complex form to a more closed object (compare also the top left outline in figure 4.7 with those below and to the right). The relationship between $DN$ and the ratio $d_h/d_I$ is shown in Figure 4.9 suggesting that $DN$ could be used to estimate the hydraulic...
diameter, if the geometric diameter is known. Further investigation of this issue is outside the scope of this thesis, and is left as a possible avenue for future research using the SLF’s extensive image archives.

Figure 4.9: Particle diameter scaling with dendricity. The ratio between hydraulic diameter \( d_h \) and mean diameter \( d_I \) as a function of dendricity \( DN \) is plotted for all experiment-averaged data.
4.3.2 Li and Pomeroy’s statistical model

The height of the wind tunnel is 1 m, and the boundary layer thickness less than that. Comparing wind tunnel data to Equation (2.18) requires the wind speed at 10 m above the surface. This is calculated from Equation 2.1 using \( u_* \) and \( z_d \), and is given in Table 4.2. The 2 m temperature is also unknown; it is assumed that in the wind tunnel there is a negligible temperature gradient normal to the surface, and hence \( T_a \) is the same as the temperature at the inlet to the test section. A comparison between Equation 2.18 and \( u (10) \) calculated from the wind tunnel threshold results is shown in Figure 4.10.

![Figure 4.10: Drift threshold wind speeds at 10 m above the surface as a function of air temperature. A 10 m wind tunnel threshold velocity is calculated from measured \( u_* \) and \( z_d \) using Equation 2.1. The solid line shows Equation 2.18 (Li and Pomeroy, 1997).](image)

Data from the wind tunnel suggests that the coefficients in Equation (2.18) could be adjusted. However, a close examination of the original data in Li and Pomeroy (1997) shows that at \(-5 < T_a < 0 \) °C, the 95% confidence interval extends over the range \( 3 < u_t (10) < 14 \) m s\(^{-1}\), and hence the wind tunnel and this prediction agree within the bounds of uncertainty. Equation (2.18) captures the lower limit to drift relatively well, and in that respect is useful in a forecasting role if the goal is to predict a possibility of drift. Also, the data set used to obtain Equation (2.18) is based on 6 years of hourly data from 16 weather stations, and so it seems unreasonable to propose a new set of coefficients based on a small number of wind tunnel data points.

4.3.3 Schmidt’s force model

Schmidt’s model was explicitly developed for the threshold of motion of ice particles, but all parameters used are tuned towards snow. To calculate \( u_* \) from Equation (2.16) requires the grain diameter \( d_I \), bond diameter \( d_b \), co-ordination number \( N_3 \) and sphericity \( SP \). Grain diameter \( d_I \) and sphericity \( SP \) are obtained from image processing of photomicrographs, while \( d_b \) and \( N_3 \) could not be obtained using the onsite characterisation, and so must be estimated.

Brown et al. (1999) assumed \( d_b = d/10 \) for fresh snow, increasing to 0.15-0.30 after 7 days of isothermal metamorphosis. In order to bracket the likely range for the snow at the wind tunnel, a range of \( 0.1 \leq d_b/d_I \leq 0.4 \) is assumed, with no correction for density or age. Brown et al. also assumed \( N_3 = 2.5 \), which from Equation (2.17) is only obtained in snow with density greater than 223 kg m\(^{-3}\). As Equation (2.17) is obtained from a fit to experiment data, we presume that it is reliable, and use that to calculate
Threshold friction velocities predicted using Equation (2.16) are shown with measured values in Figure 4.11. Results using both the original coefficients and the modified values in SNOWPACK 9.1 are shown.

![Comparison of predicted and measured u*. (a) Schmidt (1980) formulation, (b) SNOWPACK 9.1 formulation. Squares show predictions assuming \( \frac{d_b}{d_I} = 2.5 \), circles assume \( \frac{d_b}{d_I} = 10 \). The dashed lines show the case where predictions and measurements match.](image)

Figure 4.11(a) shows that \( u_* \) is significantly over-estimated by Equation (2.16) when using Schmidt’s original values for \( A \) and \( B \). The points which best fit the data are those where it was assumed \( d_b = d_I/10 \), and so the bond strength is predicted to be lowest. This partly supports the data presented in Figure 4.8, where it was seen that new snow has a similar \( A \) to dry sand; new snow typically has a lower \( N_3 \) than older snow and hence is more weakly bonded to the surface.

Figure 4.11(b) shows the calculated \( u_* \) if the coefficients in Equation (2.16) are set to \( A = 0.023 \) and \( B = 3.45 \times 10^{-3} \). These constants are implemented in SNOWPACK 9.1. The measured and calculated values are now closer, and the uncertainty is reduced. However, uncertainty remains as \( d_b \) is unknown. From Equation (2.16), \( d_b \) and \( d_I \) of the entrained grains are more important for calculating the threshold shear than the sphericity \( S_p \).

Using mean values of particle size and shape ignores the observed distributions in these parameters. Figure 4.12 shows the values of \( u_* \) that are obtained when mean and individual grain characteristics are used in Equation (2.16). The size of the bond between particles is still unknown, and so two extreme values (\( d_I = 10d_b \) and \( d_I = 2.5d_b \)) are used to bracket the prediction. The observed \( u_* \) is also shown for comparison. As the bond size would actually also follow a probability distribution function, the discrete values in Figure 4.12 could also be replaced by a probability distribution, if \( P(d_b/d_I) \) were known.
4.4 Grain sizes of drifting snow

The grain sizes of drifting snow in the wind tunnel are measured by the SPC. The influence of increasing $u_*$ can also be seen from the data. Figure 4.13 shows the same data as Figure 3.18, separated between the plateaux in the experiment. The effect of increasing $u_*$ is to increase the size of particles that can be transported under steady conditions, and also the number and frequency of particles.

Figure 4.13: Drifting particle size distribution in the measurement plateau. The upper plot shows the lowest $u_*$, the lower plot, the highest $u_*$. 'obs' is the total number of observations of drifting snow. The line is the fit to each data set using the gamma fit, Equation (3.3). Data is from the test on 9th March 2005.
Table 4.3: Gamma fit parameters for data plotted in Figure 4.13. The mean particle diameter $\bar{d} = \alpha \beta$.

<table>
<thead>
<tr>
<th>$u_*$</th>
<th>$\alpha$</th>
<th>$\beta$</th>
<th>$\bar{d} (\mu m)$</th>
</tr>
</thead>
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<tr>
<td>0.34</td>
<td>7.50</td>
<td>12.72</td>
<td>95.44</td>
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</tr>
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<td>4.51</td>
<td>29.96</td>
<td>134.97</td>
</tr>
<tr>
<td>all data</td>
<td>4.53</td>
<td>29.95</td>
<td>135.6</td>
</tr>
</tbody>
</table>

The snow used in the experiments was carefully characterised. The relationship between the snow particle diameter on the surface and the mean value found at the SPC ($d_{SPC}$) just after drift starts is shown in Figure 4.14. To use a consistent definition of diameter when comparing SPC and image processing results, the surface particle diameter is calculated as that of a circle with the same projected area $a$ as that seen on the surface. This equivalent diameter is given by

$$d_{I,eq} = \sqrt{\frac{4a}{\pi}}.$$  \hspace{1cm} (4.8)

Figure 4.14: A comparison of the diameters of airborne and surface snow particles. Airborne particle diameters are taken from a fit to data (Equation 3.3) measured at 50 mm above the surface by the SPC immediately after the start of drift. (a) Airborne particle diameters compared to surface particle equivalent diameters from images, $d_{I,eq}$. (b) Airborne particle diameters as a function of $u_*$.

Figure 4.14 shows that the size of drifting snow particles correlates well with the surface grain size, and $u_*$ during the plateau. Drifting snow particles at 50 mm above the snow surface are approximately 10% of the diameter of those found on the surface. The coefficient of correlation is $R^2 = 0.51$, and is significant at the 99% level. This dependency suggests that particles are either rapidly fracturing before they reach the detector, or that smaller ice particles are selectively entrained from the snow surface. Larger particles at the SPC are also associated with higher $u_*$, as also shown in Figure 4.3. A simple multiple linear regression of $d_{SPC}$ as $f(d_{I,eq}, u_*)$ shows that $u_*$ and the surface particle diameter account for more than 60% of the variation in $d_{SPC}$.

Simulations using the model of Doorschot et al. (2004) show that the jump length of a 150 $\mu m$ diameter particle, reaching 50 mm above the surface of fresh snow, is in the order of 1 m. If particles
are rebounding from the surface on every impact, the particles are likely to have had 3 jumps before reaching the detector. This number of impacts may be enough to reduce the size of the particles seen during the characterisation to the size of particles at the SPC, giving the relationship seen in Figure 4.13 but we cannot track individual particles through the tunnel to investigate this. For low density snow, as was often measured, and at velocities only slightly higher than the threshold, aerodynamic entrainment might be expected to dominate over rebound (Doorschot and Lehnig, 2002). Because smaller grains have smaller bonds to their neighbours and lower mass, and are therefore easier to entrain from the surface, it is most likely that it is the smaller grains which move at the drift threshold.

Drifting snow particle sizes measured in field campaigns in Antarctica reported by Budd et al. (1966) and Nishimura and Nemoto (2005), and over flat terrain in North America by Schmidt (1982) are summarised in Table 4.4. They typically measured drifting particles with diameters in the range 100-200 µm, using similar equipment to that used here, although they did not relate these back to a characterisation of the snow on the ground.

<table>
<thead>
<tr>
<th>Source</th>
<th>location</th>
<th>height (cm)</th>
<th>(d) (mm)</th>
<th>(\alpha)</th>
<th>(\beta)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Budd et al. (1966, Table 5)</td>
<td>Byrd Station, Antarctica</td>
<td>3.125</td>
<td>0.165</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>12.5</td>
<td>0.099</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>50</td>
<td>0.096</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>100</td>
<td>0.086</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Schmidt (1982, Figure 6)</td>
<td>S.E. Wyoming, USA</td>
<td>5</td>
<td>0.181</td>
<td>5.1</td>
<td>35.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10</td>
<td>0.173</td>
<td>5.3</td>
<td>32.64</td>
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<tr>
<td></td>
<td></td>
<td>20</td>
<td>0.136</td>
<td>5.3</td>
<td>25.66</td>
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<td></td>
<td>35</td>
<td>0.121</td>
<td>8.4</td>
<td>14.41</td>
</tr>
<tr>
<td></td>
<td></td>
<td>50</td>
<td>0.115</td>
<td>10.2</td>
<td>11.27</td>
</tr>
<tr>
<td></td>
<td></td>
<td>100</td>
<td>0.087</td>
<td>14.0</td>
<td>6.21</td>
</tr>
<tr>
<td>Nishimura and Nemoto (2005, Figure 7) and</td>
<td>Mizuho Station, Antarctica</td>
<td>2</td>
<td>0.170</td>
<td>5.02</td>
<td>33.77</td>
</tr>
<tr>
<td>personal communication from K. Nishimura</td>
<td></td>
<td>4</td>
<td>0.154</td>
<td>5.06</td>
<td>30.38</td>
</tr>
<tr>
<td></td>
<td></td>
<td>6</td>
<td>0.139</td>
<td>4.79</td>
<td>28.95</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10</td>
<td>0.126</td>
<td>4.76</td>
<td>26.48</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15</td>
<td>0.117</td>
<td>5.09</td>
<td>23.05</td>
</tr>
<tr>
<td></td>
<td></td>
<td>20</td>
<td>0.102</td>
<td>7.0</td>
<td>14.58</td>
</tr>
</tbody>
</table>

Table 4.4: Data from literature for drifting snow particle size in the lowest meter of the atmospheric boundary layer. Gamma coefficients \(\alpha\) and \(\beta\) are given for fits to \(d\) in µm. Interpolated or calculated data is in **bold** font.

Particle size distribution data from the all particles seen at the SPC was fitted using the gamma probability function, equation (3.3). To avoid introducing errors from large particles being truncated to the measuring range of the SPC, the data was limited to the lowest 31 of the 32 size classes that the SPC can measure. Gamma fit parameters \(\alpha\) and \(\beta\) for each test are plotted in Figure 4.15(a) and data from two tests with extreme distributions are given in Figure 4.15(b). The mean diameter detected by the SPC is in the range 50-200 µm, although the form of the distribution varies considerably. There is a trend towards decreasing mean size as the gamma fit parameter \(\alpha\) increases. Increasing \(\alpha\) implies a narrower size distribution; the trend may be a result of reduced sensor sensitivity, or thresholding of diameters, at lower diameters.

Despite the tunnel fetch of only 3 m, compared to several hundred or thousand metres in ideal field experiments, results from the wind tunnel are similar to those seen in literature. The range of the gamma fit parameters \(\alpha\) and \(\beta\) is slightly larger than those seen in literature. In particular, \(\alpha\) tends to be large when small sizes are observed, indicating distributions which have well defined peaks. This could be due to there being rather more wind tunnel measurements than field observations, or may be due to the larger variation in source snow in the wind tunnel than in the field. Repeating these measurements at different locations along the wind tunnel would show if this difference in particle size distributions was variation due to fetch or a result of the reduced mass flux in the wind tunnel, compared to field measurements.
4.5 Snow surface roughness lengths

The roughness length $z_0$ of the snow surface was calculated from the MiniAir velocity profile. Results from all tests for the threshold roughness length ($z_{0t}$) are given in Table 4.2. Measurements were also taken before the threshold, and all data are shown in Figure 4.16(a).

The threshold roughness length $z_{0t}$ measured in the wind tunnel is typically $1 \times 10^{-4}$ m, and the
4.5. SNOW SURFACE ROUGHNESS LENGTHS

average value for C is 0.014. The roughness lengths observed are lower than most other measurements of $z_0$ over snow. It is two orders of magnitude lower than those measured over fresh snow by Doorschot et al. (2004) in an alpine landscape and almost two orders of magnitude lower than at Ice Station Weddell by Andreas et al. (2005). It is an order of magnitude lower than $z_0$ measured during the Surface Heat Budget of the Arctic Ocean field experiment (SHEBA), Andreas et al. (2004). Bintanja and van den Broeke (1995) measured $z_0$ typically 1-2 orders of magnitude higher than the threshold levels in the tunnel. Figure 3 in Bintanja and van den Broeke (1995) shows a lower limit to their data at $z_0 = 0.016 u_*^2 / g$. Note that they used Charnock’s relationship for breaking waves, and so using instead the form in Equation (2.20), this gives $C = 0.032$, and a lower limit to $z_0$ which is 50% higher than these measurements. Pomeroy and Gray (1990) found $C = 0.1203$, giving $z_0$ some 6 times greater than measured in the tunnel. Higher values in natural settings compared to the wind tunnel are probably a result of the experimental set-up; the tunnel floor has less variation in height than natural flat surfaces, where sastrugi, vegetation or other obstructions would slightly increase $z_0$, or where topological effects would significantly increase $z_0$.

Figure 4.16(b) shows the measured $u_*$ and $z_0$ normalised by the threshold conditions. A linear regression fitted by least squares to the data shows that $z_0$ increases slightly as $u_*$ increases towards the threshold value, so that $z_0/ z_{0t} = 0.1 u_*/ u_{st} + 0.91$. The regression is heavily influenced by one test with a comparatively high $u_*$ (17th Jan 2005). Removing that test data gives $z_0/ z_{0t} = -0.33 u_*/ u_{st} + 1.32$. The 95% confidence interval for both cases includes the line $z_0/ z_{0t} = 1$, suggesting that it is highly likely that the $z_0$ for a particular snow surface is probably independent of $u_*$ until the start of drift. This independence from $u_*$ suggests that the surface is aerodynamically rough, and that no homogenous viscous sub-layer exists over the surface. The wall thickness $\delta_w$ was estimated to be about 50 $\mu$m in these conditions (see Section 2.1.1), giving a limit to the viscous sublayer of about 250 $\mu$m. That distance is less than the usual height variation of the surface, and similar to the roughness length of the surface, which further indicate a hydraulically rough surface. This may be linked to through flow in the snow surface, as discussed in Chapter 5.

Algorithms developed for the drift threshold shear for sand can be applied for snow by converting mean diameters to the hydraulic diameter and including cohesive effects. By implication, the same approach should be applicable for the roughness length, which over flat granular surfaces is usually taken to be a linear function of the particle diameter (see Table 2.2). Roughness lengths before drift should be independent of cohesion, and so algorithms for sand should be directly transferable, giving $d/30 \leq z_0 \leq d/8$. Results from experiments are plotted in Figure 4.17 along with lines of $d/n$, where $n$ is some constant. The particle diameters $d$ that are used are $d_t$, the mean value from images; $d_{t,95%}$, the diameter which only 5% of particles are larger than and $d_h$, the hydraulic diameter. Figure 4.17 shows that there are no clear relationships between any diameter measures and $z_{0t}$. The highest coefficient of determination found was $R^2 = 0.23$ between $d_{t,95%}$ and $z_{0t}$.

This result should be compared to the data for the threshold friction velocity and roughness lengths. Substituting the equation for the threshold friction velocity $u_{st}$ (Equation 2.13) into Owen’s boundary layer modification equation (see Equation 2.20) gives the following relationship:

$$z_{0t} = \frac{C}{2g} A^2 \frac{\rho_{ice} - \rho_{air}}{\rho_{air}} d_h.$$  \hspace{1cm} (4.9)

Taking $C = 0.021$, $A = 0.18$ and typical values of $\rho_{ice} = 917 \text{ kg m}^{-3}$ and $\rho_{air} = 1.225 \text{ kg m}^{-3}$ this gives $z_{0t} \approx d_h/4$. There is a strong trend towards this in Figure 4.17 except for the data from the depth hoar experiments of 17th January 2005. This value implies that snow surfaces have roughness lengths some 7.5 times those of sand or other flat, granular material of the same diameter.

The roughness length to grain size correlations for sand are based on experiments over flat, smooth and uniform beds of sand. In those experiments, the sand grains define the topology of the surface; there is no larger-scale deviation of the surface. By comparison, surface features are often seen in snow, ranging from ‘dimples’ with scales of 10 mm or less, to waves or dunes with scales of around a meter. These small-scale features were seen on the snow surfaces used in the wind tunnel before experiments, although it was not possible to characterise them. It is likely that these features influence the surface roughness more than the grain, and so some effort should be made to quantify this in future. A good example of this is given by Lancaster et al. (1991), who found that $z_0$ scaled with the root-mean-square surface feature size over a desert surface.
4.6 Roughness lengths with drift

Friction velocities and roughness lengths were measured at $u_\ast > u_\ast t$ using the dynamic pressure rake, as the rake is less sensitive to impacting drifting snow and has increased vertical resolution, compared to the miniairs. This was only possible during Winter 2005/2006, when both natural snow and sieved snow were used. Information about the weather and snow characterisation is given in Tables 3.5 and 4.1.

All $u_\ast$ and $z_0$ measurements from the dynamic pressure rake are shown in Figure 4.18. Data is categorised by the mean frequency of drift detected at the SPC during a measurement plateau, $P(\text{drift})$. Data from the same experiment are connected by lines. When drift was detected at the SPC, data used to calculate $u_\ast$ and $z_0$ was limited to $z > 50$ mm. The relationship $z_0 = 0.021 u_\ast^2/2g$ is also plotted, along with error bands of $\pm 25\%$. Results in plateaux where drift occurred more than 25% of the time tend towards $C = 0.021$. There is some indication that $z_0$ adjusts towards the value $0.021 u_\ast^2/2g$ from pre-threshold levels, but there are also experiments where $z_0$ stays consistently at much lower values than would be expected.

An adjustment towards a constant value of $C$ with increasing drift frequency is a logical result of drift over a surface with a low roughness length. As drift becomes sustained and regular ($P(\text{drift}) \to 1$), $z_0$ over generally smooth surfaces is determined by the drift, and the amount of drift to do this will vary, depending on the surface topology. Where the surface topology is larger than the drift height, the drift will not influence the roughness, regardless of the frequency of transport. However, in all of the cases presented here, drift was detected at 50 mm above the surface, reaching a larger height than any observed topology, and hence the drift would be expected to set $z_0 \propto u_\ast^2$, giving a constant value of $C$ in each experiment.

Figure 4.18 suggests that it is quite probable that snow drift has a similar influence on the roughness length of a surface, to sand or earth, as suggested by Owen (1964). Intuitively, this makes sense; drifting snow, sand and earth are all basically drifting granular materials, and might be expected to have similar influences on the form of the boundary layer.

A goal of this thesis was to assess the shear within the saltation layer. Velocity profile measurements within the saltation layer are influenced by the drifting material and the profiles are noisy (see for example Figure 3.10). Fitting the profiles at heights below the SPC with the standard log profile (Equation...
Figure 4.18: Owen’s $C$ as a function of drift intermittency. $P(drift) = 0$ is the case with no drift, $P(drift) = 1$ is the case with drift detected continuously during the measurement plateau. Data from the same experiment are connected by lines and plateaux are indicated by ◦. The line $C = 0.021$ is also plotted and error bars at $C \pm 25\%$ are shown as a guide to uncertainty.

often resulted in fits where the correlation coefficient between the fit velocity and measured velocity was less than 0.5, compared to more than 0.95 for profiles above the saltating material. It is also known that the total shear stress in the boundary layer stays constant when particles are in motion, and so drifting particles result in a drop of airborne shear (Raupach, 1991). The shear varies with height, and so the shear cannot be quantified by fits to the velocity profile in the drifting cloud using the log law, which assumes a constant shear in the region where it is applied. Other measurement techniques, such as hot-wire anemometry using heavy-duty film or wedge-shaped probes giving point measurements of the shear just above the surface, or surface drag plates as in Nemoto and Nishimura (2001), may be more useful for answering this question than mean-velocity profile measurements.

### 4.7 Summary

Boundary layers and mass transport rates over snow have been measured during a series of 22 experiments with smooth, fresh snow surfaces. Snow characteristics were measured at the grain and bulk scales. A routine has been developed to automatically recognise the drift threshold from boundary layer data and point measurements of drifting snow mass flux. The threshold $u_*$ were found to vary between 0.27 and 0.69 m s$^{-1}$. Roughness lengths at the threshold varied from 0.04 to 0.13 mm, and the range of both $u_*$ and $z_0$ may increase as more snow samples are tested. The roughness length increases slightly before the onset of drift, but the variation is within the range of measurement uncertainty. A constant roughness length would be expected for a rough surface.

Threshold algorithms based on force balance (Bagnold, 1941) show that the drift threshold for new and decomposing snow is well predicted using the particle hydraulic diameter and a threshold parameter $\mathcal{A} \approx 0.18$. This value is 80 % greater than found for non-cohesive, granular material such as sand. The threshold friction velocity predicted by Schmidt’s algorithm using mean particle shape information from photomicrographs was found to be highly dependent on the assumed bond size. If large bonds were assumed, predicted $u_*$ were $\approx$100% larger than the observed mean value in the wind tunnel. A modified formulation as used in SNOWPACK 9.1 also over-predicted the threshold by $\approx 50\%$. How-
ever, assuming small bond sizes in both formulations resulted in good agreement. Models based on Schmidt's algorithm can bound $u_{*t}$ only if the co-ordination number, particle and bond diameter are accurately described by models which reflect the variation found in a snow pack.

Grain sizes measured by an SPC 50 mm above the snow surface directly after the start of drift were fitted using a gamma distribution. This gave a mean particle diameter in the range 75-200 microns, and most of the variation in particle size was explained by variation in the surface particle size and $u_{*t}$. The airborne particle diameter is similar to that found in field experiments in both Antarctica and Wyoming, although with occasional narrower size ranges. The size distribution of the transported particles is heavily influenced by $u_{*t}$.

Wind-tunnel $z_0$ are smaller than most field measurements, although similar to the lowest values measured over Antarctic snow-covered ice with minimal topographical influences. The value of $z_0$ was found to be almost constant as $u_s$ increased toward $u_{*t}$, as has been observed for other rough surfaces. Simple $z_0 = f(d)$ correlations developed for sand under predict the roughness length of snow of the same hydraulic diameter. The roughness length at the drift threshold was well predicted using the approach of Owen (1964), with $C = 0.021$, as has been found for sand and soil. Measurements using the dynamic pressure rake, which measures above the saltating cloud, show that the constant of proportionality $C$ seems to converge on $C = 0.021$ as the frequency of drift increases and comes to dominate the surface roughness.
Chapter 5

Snow as a porous medium

Twinkle, twinkle little star,
How I wonder what you are?

Traditional

Work in the wind tunnel is to some extent limited by the choice of snow as a research topic. For more than 8 months of the year, mean daytime temperatures at the wind tunnel rise above freezing, preventing work with snow. Also, long-duration measurements are difficult because of metamorphosis of the snow surface and erosion.

As was seen in the previous section, measurements over snow show that $z_0$ before the onset of drift is probably independent of $u_*$ (see Figure 4.16(a)), and that $z_0$ best correlates with the bulk parameters of the snow (see Figure 4.3). The roughness length does not seem to correlate well with the grain size (see Figure 4.17), and this observation and Figure 5.2 suggest that the snow may also be well described using parameters which relate to the foam permeability, porosity and bulk structure. Some investigators have also noted that permeable surfaces have increased roughness compared to impermeable surfaces with the same geometry (e.g. Zagni and Smith, 1976; Zippe and Graf, 1983; Nikora et al., 2002; Breugem et al., 2006), which may explain these results. Together, these are good reasons to investigate the role of permeability of flat surfaces in raising $z_0$.

A further motivation comes from wanting to link the length scales which describe the bulk properties of the material with the boundary layer characteristics. These length scales include the length scales of the intact material, rather than just those associated with the individual particles, and require some understanding of the 3-dimensional form of snow. Measurements of the intact material proved extremely difficult, if at all possible, in the wind tunnel. By comparison an artificial material has the advantage of being more stable, homogenous and can be delivered to suitable measurement equipment without disintegrating or melting.

For these reasons measurements were also made over an artificial foam with the aim of helping to explain results from snow. A basic overview is given of the microstructure of snow and compared to the foam. The boundary layer over the foam is measured as the depth, structure and layering of the material changes. The results of these experiments illustrate the influence that the different parameters of the porous medium have on the overlying boundary layer, and the possible implications for snow surfaces are discussed. Much of this chapter appeared in Clifton et al. (2007).

5.1 The microstructure of snow

Snow has traditionally been seen as a regular, 2-dimensional material; Kepler waxed lyrical in 1661 about its symmetry about a central axis, and children have for years been entertained by folding paper and then cutting out sections to make ‘snow flakes’, occasionally with the correct 60° symmetry. Both of these examples serve to emphasise the perceived ‘flat’ nature of fresh snow. Photos taken of surface snow (Figure 5.1) also reinforce this idea of clearly visible, discrete crystals. Similarly, the fact that ‘particles’ are detected in the air, and the geometry of individual crystals is described by the morphology diagram (Figure 4.5), may subconsciously encourage this idea.

Whilst this flatness and discrete granular form might be accurate for fresh snow fall direct from a cloud (Magono and Lee, 1966), as soon as snow lies on a surface and is then exposed to temperature or
water vapour gradients, a snow flake loses its individuality and merges into the whole surrounding it, forming an ice matrix filled with air. The same is true even after a few minutes; whilst individual crystals might apparently be visible to the naked eye or through a lens, closer inspection shows that these crystals have connected at one or more points to their neighbours, and have formed a continuous structure. Figure 5.2 shows a 3-dimensional reconstruction of the structure of a sample of fresh snow, obtained with the SLF’s X-ray computer tomograph (CT scanner, Scanco Medical AG model μCT 80, see Schneebeli and Sokratov, 2004). The CT scanner allows non-destructive measurement of the structure with 10 micron resolution. An open-celled polyester foam is shown for comparison.

Figure 5.2: Three-dimensional reconstruction of the microstructure of foams. (a) New snow, density 54 kg m\(^{-3}\). The scale bar is 1 mm. (b) Regicell 60 reticulated foam. Material structure was imaged using the SLF CT scanner, and the scale is identical for both images.

The 3-dimensional form of the snow is clear, and the individual crystals and forms which are so distinct in Figure 5.1 are virtually invisible in Figure 5.2. Also, snow and reticulated foams clearly share geometric characteristics, and both are highly porous and have regular structures. Snow is known to
be isotropic over small vertical distances \cite{Colbeck1991}, permeable \cite{Colbeck1989} and composed of identifiable grains and bonds or chains of material \cite{Gubler1978}. High-quality foams are also isotropic and available as open celled media. In open-celled foams the structures formed by the material are characterised by small nodes connected by branches \cite{Gibson1988}. The porosity of foams and new snow is similar, at more than 90%.

To investigate the influence of bulk structure on the boundary layer that forms over a porous media, measurements were also made over 3 different types of polyester filter foams from Fritz Nauer AG, Wolfhausen, Switzerland. The foam characteristics are detailed in Table \ref{table:foam_properties}. The given names are trade names. As there is no granular structure to the foams, unlike in the snow, other geometry is used to characterise the foam. This is obtained by 3-dimensional imaging, using the SLF X-ray tomograph. The data which is obtained includes the porosity $\phi$ (percentage of the sampling volume which is air), the pore length $\ell_{\text{pore}}$, the node size $\ell_{\text{node}}$ (the mean size of the solid elements) and the specific surface area SSA, defined as the surface area of the matrix per sample volume. While the porosity stays almost constant, the other values change between the different foams by about an order of magnitude.

Table \ref{table:foam_properties}: Geometric characteristics of foam and snow. Foam and new snow (as shown in Figure \ref{fig:new_snow}) data was measured by X-ray tomography. Other snow data, given by experiment date, uses the thin-section stereological method of \cite{Baddely1986}.

<table>
<thead>
<tr>
<th>Material</th>
<th>Porosity ($%$)</th>
<th>$\ell_{\text{pore}}$ (mm)</th>
<th>$\ell_{\text{node}}$ (mm)</th>
<th>SSA (mm$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>new snow (see Fig. 5.1)</td>
<td>94.1</td>
<td>0.46</td>
<td>0.06</td>
<td>2.88</td>
</tr>
<tr>
<td>natural snow, 21/02/2006</td>
<td>89.1</td>
<td></td>
<td></td>
<td>4.97</td>
</tr>
<tr>
<td>natural snow, 28/02/2006</td>
<td>93.1</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>natural snow, 2/03/2006</td>
<td>92.0</td>
<td></td>
<td></td>
<td>4.49</td>
</tr>
<tr>
<td>sieved snow, 14/03/2006</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>sieved snow, 21/03/2006</td>
<td>53.8</td>
<td></td>
<td></td>
<td>4.47</td>
</tr>
<tr>
<td>sieved snow, 22/03/2006</td>
<td>56.3</td>
<td></td>
<td></td>
<td>4.01</td>
</tr>
<tr>
<td>sieved snow, 23/03/2006</td>
<td>43.1</td>
<td></td>
<td></td>
<td>4.13</td>
</tr>
<tr>
<td>Regicell 10</td>
<td>96.4</td>
<td>3.89</td>
<td>0.27</td>
<td>0.37</td>
</tr>
<tr>
<td>Regicell 30</td>
<td>97.0</td>
<td>1.47</td>
<td>0.11</td>
<td>0.82</td>
</tr>
<tr>
<td>Regicell 60</td>
<td>98.0</td>
<td>0.49</td>
<td>0.03</td>
<td>1.70</td>
</tr>
</tbody>
</table>

Table \ref{table:foam_properties} also gives the foam permeability, $k$. Permeability is a measure of momentum lost by a fluid as it passes through a material, and was measured by T. Neumann (U. Vermont) and Z. Courville (U. Dartmouth) using a through-flow permeameter \cite{Albert2000}. Results are shown along with literature data in Table \ref{table:permeability}. The permeability of snow is similar to that of the lowest permeability foam. The length scale $1/SSA$ correlates highly with the square root of the permeability of the foam.
Table 5.2: Porosity and permeability ($k$) of different snow types from literature, compared to foams.

<table>
<thead>
<tr>
<th>Material</th>
<th>Porosity (%)</th>
<th>$k$ ($\times 10^{-9}$ m$^2$)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>wind-blown snow</td>
<td>83-85</td>
<td>2.6-3.1</td>
<td>Seasonal snow cover (Sommerfeld and Rocchio, 1993)</td>
</tr>
<tr>
<td>wind-packed snow</td>
<td>77</td>
<td>1</td>
<td>Antarctica (Albert et al., 2000)</td>
</tr>
<tr>
<td>firn</td>
<td>-</td>
<td>28</td>
<td></td>
</tr>
<tr>
<td>melt-freeze surface</td>
<td>63</td>
<td>6-14</td>
<td>Seasonal snow cover (Albert and Perron, 2000)</td>
</tr>
<tr>
<td>melt-freeze crystals</td>
<td>62</td>
<td>5-11</td>
<td></td>
</tr>
<tr>
<td>ice layer</td>
<td>-</td>
<td>0.03-1.4</td>
<td></td>
</tr>
<tr>
<td>Regicell 10</td>
<td>96</td>
<td>160</td>
<td>T. Neumann and Z. Courville (personal note, 2006)</td>
</tr>
<tr>
<td>Regicell 30</td>
<td>97</td>
<td>39</td>
<td></td>
</tr>
<tr>
<td>Regicell 60</td>
<td>98</td>
<td>6</td>
<td></td>
</tr>
</tbody>
</table>

5.2 Boundary layers over porous media

Boundary layer profiles were measured over the materials described in Table 5.1 using the dynamic pressure rake. The pitot rake was positioned 2.4 m downstream of the start of the media, and 5 to 6 measurements were made between 5 mm above the surface and the edge of the logarithmic region, which was usually around 10 cm above the media. First measurements were made over snow and then continued over foam and a solid surface. The wind tunnel set-up is shown in Figure 5.3.

![Figure 5.3: Schematic view of the wind tunnel experiment. The porous media used in this investigation were snow, and open-celled polyester foams. The SPC was only present during experiments with snow.](image)

Before the measurements over foam and snow, the static pressure gradient along the tunnel was set as close as possible to zero over a wooden floor in the test section by flaring the tunnel roof to compensate for the boundary layer growth. The roof was then kept in the same position for experiments with different surfaces in the test section. The foams were also of different depths, ranging from 20 to 150 mm thickness. The upper surface of the foam was level with the floor upstream of the test section. Static pressures measured at 0.5 m intervals along the wind tunnel are shown normalised by the free-stream dynamic pressure in Figure 5.4. No detailed pressure gradient data is available for the snow surfaces because of the length of time required to measure the pressure profile. Instead, inclined manometers were connected to static pressure taps on the tunnel wall during tests with snow and checked manually. The pressure gradients over all surfaces in the wind tunnel test section were less than 2% of the reference dynamic pressure, per meter.

The influence of the different floor coverings in the test section ($x > 8$ m) is clear. Adding the foam in place of the smooth wooden floor results in an increase in pressure drop in the wind tunnel, suggesting a change in drag in the test section. The increase is greatest in the upstream part of the test section, and pressure gradients are low around the rake, and so the wind tunnel flow in the region of the rake is considered practically zero pressure gradient. Creep velocities introduced as a result of the small pressure gradient are estimated to be around 1 mm s$^{-1}$. As would be expected, the different surfaces
5.2. BOUNDARY LAYERS OVER POROUS MEDIA

Figure 5.4: Static pressure difference along the wind tunnel. Data is the difference in pressure to a point 0.25 m downstream of the spires, normalised by the reference dynamic pressure. The test section floor (x > 8 m) is replaced by wooden boards or foam. Data shown are means for 5 different dynamic pressures with each different floor. The legend details the foam type ('Regicell nn',) and depth ('zz mm').

in the test section do not alter the flow upstream. The observed increase in pressure drop as the surface becomes more permeable is consistent with results from numerical simulations (Jiménez et al., 2001; Breugem et al., 2006), which show a 10-30% increase in duct friction factor when a solid wall is replaced with a permeable wall.

Breugem et al. (2006) reports a reduction of the Von Kármán constant, κ, in the presence of a permeable wall. This modification was seen in a numerical simulation of flow in a duct with a permeable wall where the permeability Reynolds number, \( \text{Re}_k = u_\ast \sqrt{k/\nu} \), was greater than 1, and calculated from the Reynolds stress profile. In those simulations κ reduced to 0.31 for \( \text{Re}_k = 1.06 \), and to 0.23 for \( \text{Re}_k = 9.35 \). In the experiments reported here, \( \text{Re}_k > 1 \), raising the possibility that the von Kármán constant is lower than the 0.41 that was assumed, and that the calculated \( u_\ast \) is higher than that which would be found from measurements of the Reynolds stresses. However, the mechanism for reducing κ is unclear, and so all measurements reported here assume that κ is unchanged. The relative paucity of literature investigating flow over porous boundaries makes this risk difficult to assess, but could be checked with direct measurements of Reynolds stress using hot-wire anemometers. The reported roughness length \( z_0 \) should however be independent of κ as it is determined from regression of \( u(z) \) as a function of \( \ln(z) \).

5.2.1 Roughness lengths of different surfaces

Boundary layer velocity profiles were measured using the dynamic rake, as described in Section 3.4. Friction velocity \( u_\ast \), roughness lengths \( z_0 \) and zero-plane displacement heights were calculated for all profiles, and in every case over foam, the coefficient of determination between the measured and fit data using Equation (2.5) was \( R^2 > 0.99 \). Measured \( u_\ast \) and \( z_0 \) over the foam and over snow surfaces (limited to measurements with no drift) are shown in Figure 5.5.

Two different lines of constant \( \text{Re}_* \) are also plotted in Figure 5.5. Raupach et al. (1991) shows that for a smooth wall, \( z_0 = \frac{\nu}{u_\ast} \exp(-\kappa C_s) \), which is shown as a shaded region in Figure 5.5 between \( C_s = 5 \) (Schlichting and Gersten, 2003) and \( C_s = 5.5 \) (Bhaganagar et al., 2004); see also Section 5.4.2. \( \text{Re}_* = 2.5 \) is also plotted, which is the lower limit to the hydraulically rough regime. In the hydraulically rough
regime, where $Re_\ast \geq 2.5$, $z_0$ is usually assumed to be independent of $u_\ast$.

Figure 5.5: $u_\ast$ and $z_0$ measured over (a) snow and (b,c) reticulated foams. Snow data are grouped by experiment date and limited to profiles where no drift was measured. Foam data are grouped according to depth and type of foam. Data is limited to velocity profiles where the fit using the log law is $R^2 \geq 0.9$. The shaded region is $0.105 \leq Re_\ast \leq 0.129$, which is the range of literature data for a smooth surface (Schlichting and Gersten, 2003; Bhaganagar et al., 2004). The line $Re_\ast = 2.5$ (--), which is the lower limit of the rough regime, is also shown.

Figure 5.5 shows that the modification of the boundary layer inferred from Figure 5.4 was correct, shown here by the large differences in $z_0$ between the different surfaces. The measurements demonstrate that the solid wooden surface is in the hydraulically smooth regime (see also Figure 3.7), and as was also shown in Figure 5.5 corresponds well with literature data.

The results show that although the foam surfaces are flat on a large scale, the roughness length changes between the different foams. Zippe and Graf (1983) also observed a similar effect, noting that
5.2. BOUNDARY LAYERS OVER POROUS MEDIA

a permeable surface had a higher \( z_0 \) than an impermeable surface composed of the same material. The most permeable surfaces have \( Re_* \) that places them well into the rough regime, while the least are in the transitional range. In the hydraulically rough regime, it is usually assumed that the roughness length \( z_0 \) is independent of \( u_* \). However, these results confirm the observation of Zagni and Smith (1976) that over a porous surface, \( z_0 \) is a weak function of \( u_* \). This is an important feature of permeable surfaces compared to solid rough surfaces. At lower permeabilities, the rise in \( z_0 \) with \( u_* \) is less marked. The measured \( z_0 \) for the surfaces compares well with that seen for permeable natural materials; for example the 60 ppi foam has a similar roughness length to the snow shown in Figure 5.5, and \( z_0 \) of the 10 ppi foam is similar to that for short grass of varying density, as summarised in Garratt (1994, Table A6, p. 290).

Repeated measurements over different depths of Regicell 10 show no large differences in \( z_0 \), but measurements over considerably different materials (wood and foam, for example) result in markedly different boundary layers. It is also seen that the depth of the foam used has no discernable influence on the roughness length, suggesting that the roughness length is a property which is influenced by the conditions in the top 25 mm (or less) of the porous media. This depth still represents a depth of at least 6 times the mean pore length scale.

Several length scales can be defined for the porous media. one is the square root of the permeability, \( \sqrt{k} \). Another is the equivalent grain diameter \( d_v \), and can be established from SSA and porosity using Equation (2.6). The change in \( z_0 \) normalised by these length scales, as a function of \( Re_* \) is plotted in Figure 5.6. These non-dimensional values are also reported by Breugem et al. (2006), and his simulation results are very similar to those seen in the wind tunnel. In all cases, the ratio \( z_0/d_v \) for foam is higher than that usually assumed for solid surfaces, as described in Table 2.2.

5.2.2 Zero-plane displacements

The zero-plane displacement of the velocity boundary layer over snow was calculated for profiles where no drift was observed by the SPC. This does not preclude some drift at lower heights, which may have influenced the velocity profiles near to the surface. Results (not plotted) show that the zero plane displacement was typically \( z_d \leq 1 \) mm, which was a similar size to the variation in height of the surface. The observed magnitude of \( z_d \) is similar to that seen by Kobayashi (1969) in wind tunnel experiments over an artificial snow surface, although no information is available about the snow permeability in Kobayashi’s experiments. Also, applying a displacement model did not significantly improve the \( R^2 \).
value of the fit, compared to a normal log-law.

The zero-plane displacement of the boundary layers over foam was also calculated. Results are plotted in Figure 5.7 and show that $z_d$ varies strongly between materials, and generally increases with $u^*$. The zero-plane displacement over foams with permeability similar to snow (Regicell 60) is apparently non-zero at low $u^*$, but becomes zero at high $u^*$. This change in behaviour could be a result of increased measurement uncertainty at low $u^*$, rather than a real displacement, given the considerably different trend to the other foams.

![Figure 5.7: Zero-plane displacement of velocity boundary layers over foam. Zero-plane displacements were calculated from velocity profiles measured using the dynamic pressure rake. The resolution limit is about 1 mm.](image)

The zero-plane displacement, normalised by the square root of the permeability and the foam equivalent grain diameter $d_v$ are shown in Figure 5.8 as a function of the permeability Reynolds number $Re_k = u^* \sqrt{k}/\nu$. Results agree well with those from the direct numerical simulations of Breugem et al. (2006), which are plotted for comparison. The simulation results show that for the low permeability foams, the ratios $z_d/d_v$ and $z_d/\sqrt{k}$ decrease, which is opposite to the trend shown by the foam. This suggests that the non-zero displacement for the lowest permeability foam is an error. Non-dimensional values are summarised in Table 5.3.

<table>
<thead>
<tr>
<th>Foam</th>
<th>$Re_k$</th>
<th>$Re_*$</th>
<th>$z_0/\sqrt{k}$</th>
<th>$z_0/d_v$</th>
<th>$z_d/\sqrt{k}$</th>
<th>$z_d/d_v$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Regicell 30</td>
<td>5 - 11</td>
<td>20 - 106</td>
<td>4 - 10</td>
<td>4 - 9</td>
<td>15 - 29</td>
<td>13 - 26</td>
</tr>
<tr>
<td>Regicell 60</td>
<td>1 - 3</td>
<td>1 - 4</td>
<td>1 - 2</td>
<td>1 - 2</td>
<td>0 - 16</td>
<td>0 - 19</td>
</tr>
</tbody>
</table>

The influence of depth

Measurements were made with 4 different depths of Regicell 10. The pressure drop in the test section is shown enlarged in Figure 5.9.
5.2. BOUNDARY LAYERS OVER POROUS MEDIA

Figure 5.8: Normalised zero-plane displacements of velocity boundary layers over foam. Data is plotted as a function of the permeability Reynolds number, $Re_k = \frac{u^*}{\sqrt{k/\nu}}$. (a) $z_d/\sqrt{k}$. (b) $z_d/d_v$. Data from Breugem et al. (2006) is plotted for comparison.

Figure 5.9: Static pressure difference along in the test section with Regicell 10. Data is limited to the test section and Regicell 10, but is otherwise identical to Figure 5.4. The legend details the foam depth ("zz mm").

The pressure measurements show that the depth of foam does not greatly influence the bulk velocity in the wind tunnel. The influence of the depth on the velocity is small compared to the influence of the different roughness along the tunnel, as would be expected for a large change in surface drag between the media. The flow within the media represented by the zero-plane displacement of at most 20 mm is small compared to the total flow in the wind tunnel, which has a depth of 1000 mm in the test section.

The influence of depth can be seen more clearly in Figure 5.8. In that plot, the zero-plane displace-
ment reduces as the material depth decreases. The limiting case for a material of finite depth is that the zero-plane displacement height tends to the material depth, \( z_d \rightarrow z_{\text{mat}} \). At the limit, increasing \( Re_u \) no longer alters \( z_d / \sqrt{k} \). For the 150 mm deep Regicell 10, this would be at \( z_d / \sqrt{k} = 375 \), and for the 100 mm deep Regicell 60 at \( z_d / \sqrt{k} = 1325 \); in these cases it might be assumed that the behaviour of the surface approximates a material of infinite depth. However, in the case of 25 mm deep Regicell 10, this condition is reached at \( z_d / \sqrt{k} = 62.5 \). There is a suggestion from Figure 5.8 that the zero-plane displacement of the thinner foams is slightly reduced at higher \( u_u \), relative to the thicker foams, as would be expected. Any trend towards a physically required asymptotic value, the material depth itself, is harder to identify. The trend for decreasing zero-plane displacement depth with decreasing material depth is consistent with the requirement for vanishing boundary layer displacement when the material depth is negligible.

5.3 Application to snow surfaces

Results from the measurements over foam, and their good agreement with the simulation data of Breugem et al. (2006) as shown in Figure 5.6 and 5.8, suggests that the roughness length and zero-plane displacement depth of a flat, porous surface can be predicted if the permeability, SSA and \( u_u \) are known. Does the same hold for snow? Permeability data is not available for the snow surfaces where \( z_0 \) and \( u_u \) is known, so the relationships seen in Figure 5.8 cannot be confirmed for snow. The snow SSA and porosity is known (see Table 5.1), so the equivalent grain diameter \( d_v \) can be calculated. The particle hydraulic diameter \( d_h \) can also be obtained from images. Therefore the relationship seen in Figure 5.8(b) between \( z_0 / d_v \) and \( Re_d \) can be tested for its applicability to snow. However, as \( z_0 \) appears on both axes, \( Re_d = u_u d_v / \nu \) is used instead. Results for the foam and snow are plotted in Figure 5.10.

Figure 5.10: Roughness lengths and length scales in the wind tunnel. Results are split between those for foam and those for snow, without drift. Roughness lengths \( z_0 \) are the data points plotted in Figure 5.9. Length scales for snow are the equivalent particle diameter \( d_v \) and the hydraulic diameter \( d_h \).

The influence of the bulk geometry is apparently much higher in the foam than in the snow. The ratio of \( z_0 / d_v \) varies clearly with \( Re_d \) in the foam, but the snow data, regardless of which length scale is used, forms a cloud of data points, showing that the surface roughness is decoupled from the bulk properties. This de-coupling was also seen in Figure 5.3 in other measurements over snow. Further evidence of a de-coupling comes from the data shown in Table 5.1 and Figure 5.5; the snow SSA hardly
changes, while the roughness length varies considerably. In comparison, the SSA of the foam varies by almost an order of magnitude, and the measured roughness lengths by almost two orders of magnitude. This suggests that the observed high influence of the foam form on the roughness length is probably coupled to the high permeability, as only high permeability allows significant flow in the upper region of the foam and thus an influence of the foam bulk parameters on the drag.

The data obtained for smooth foam surfaces can be applied to snow by analogy. In flows over snow, \( u_* < 1 \text{ m s}^{-1} \), and \( z_0 < 1 \times 10^{-2} \text{ m} \); see the maximum values in the data of [Andreas et al. (2005), or data in Table 2.1]. This corresponds to wind speeds of less than 7.5 m s\(^{-1}\) at 2 m above ground, and up to 10 m s\(^{-1}\) at 10 m height, as shown in Figure 2.3. These velocities are frequently measured in Antarctic regions dominated by katabatic winds (Bintanja, 2001b).

The permeability of snow, in particular surface snow which has been exposed to drift, is low at \( k < 3 \times 10^{-9} \text{ m}^2 \), less than that of the least permeable foam. The zero-plane displacement of that foam is less than 1 mm, and frequently zero, and if the snow surface were flat, it might then be expected that no displacement would be found over snow in a zero pressure gradient. A flat snow surface is therefore effectively impermeable to shear-driven ventilation.

There is however an indication from literature (Kobayashi, 1969; Colbeck, 1989; Sturm and Johnson, 1991; Sokratov and Sato, 2001, 2003; Albert and Schultz, 2002; Albert et al., 2002) that some flow can be found within the snow pack and that by implication displacement heights are not always zero. In that case, it is more likely that the flow is induced by local pressure gradients. Local pressure gradients could be caused by flow around surface topology on a small scale, such as grain clusters or sastrugi, leading to ventilation of the upper few millimetres of the snow. Larger scale features such as hills or ridges create pressure gradients, and thus ventilation, over meter to hundred meter scales (Waddington and Cunningham, 1996).

5.4 Summary

A series of boundary layer velocity profile measurements were made over deep foam beds in a zero or low pressure gradient. These included measurements over different depths of foams, and a smooth solid floor. Foams were characterised in terms of permeability and geometric characteristics, as was the snow. The data collected over the flat, smooth and uniform foam has clear implications for similar measurements over snow surfaces. This includes;

**Wind tunnel response to different surfaces.** The wind tunnel boundary layer clearly responds to changes in the surface in the test section. This is seen in the pressure gradient through the tunnel and also in the parameters which describe the boundary layer, \( u_* \) and \( z_0 \). The velocity boundary layer over a solid floor matched the universal law of the wall.

**Strong influence of permeability on \( z_0 \).** A flat, permeable surface has a considerably higher roughness length than a flat impermeable surface. The roughness length increases as surface permeability rises.

**Low influence of \( u_* \) on \( z_0 \).** The measurements over foam show that in stable, non-accelerating flow, \( z_0 \) increases slightly with \( u_* \), and that this correlates well with the displacement height, \( z_d \). When the permeability is low, the displacement height tends to zero and the roughness lengths do not change with \( u_* \). This strengthens the assumption made from Figure 5.16(b) that \( z_0 \) of a snow surface (which is of low permeability, compared to the foams) is constant before the start of drift.

**Influence of material depth.** \( z_0 \) is independent of the material depth while the zero-plane displacement is smaller than the material depth. The only observed influence of material depth was on the zero-plane displacement, which decreases slightly as the material depth decreases. This effect may change if the material depth decreases further and the zero-plane displacement is constrained.

**Hydraulic regimes.** No indication was seen of a ‘smooth’ hydraulic region over the foams or snow, where \( z_0 = f (1/u_*) \), as has been suggested by e.g. Bintanja and van den Broeke (1995) or Andreas et al. (2005). This implies that the regular surface roughness elements, followed by voids in the surface, serve to break up any laminar boundary-layer region. This may be due to the presence of
significant-sized voids at the surface, which thus allow turbulent eddies to extend to the wall and corresponding losses from wake formation within the media.

**Links to material length scales.** It was seen in Figure 5.5 that the roughness length of foams increased slightly with $u_*$. Also, the zero-plane displacement and roughness length for each foam correlates at $R^2 \geq 0.97$. This suggests that it is the displacement of the boundary layer into the material which increases the effective drag of the surface, raising $z_0$. Zippe and Graf (1983) observed a similar effect, with a permeable surface having a higher $z_0$ than an impermeable surface composed of the same material.

A similar, slight increase in $z_0$ with $u_*$ is seen over snow (see Figure 4.16(b)). However, as the permeability of snow and hence $z_d$ is much less than that of foam, this is unlikely to be caused by ventilation. Instead, the increased roughness of snow compared to similar sized spheres suggests that the surface topology may be important for roughness, where increasing $u_*$ may result in increased losses around the surface structures, and thus increased $z_0$. 
Chapter 6

Modelling wind tunnel snow transport

Everything should be made as simple as possible, but not simpler.

Albert Einstein (1879-1955)

This chapter describes modelling investigations of drifting snow in the SLF wind tunnel. Several experiments are chosen for modelling using a saltation model. The numerical stability of the model is checked, and an approach to determining vertical mass flux profiles is discussed. Results from the model are compared to experiment data and field observations of drifting snow. Difficulties and potential solutions for modelling wind tunnel drifting snow are identified. A large part of the material in this chapter appears in Clifton and Lehning (in preparation).

6.1 Drifting snow in the wind tunnel

As was noted in Section 2.2, drifting and blowing snow can be separated into several types of motion, depending on the paths that the snow particles follow. Particles can simply roll along the ground in reptation, follow parabolic trajectories through the air in saltation, or be transported as a tracer in suspension.

Particles begin to move in reptation when the forces acting on them (both from aerodynamic effects and impact of drifting particles on the surface) are sufficient to break bonds and free them from the surface, and the aerodynamic drag is enough to transport them along the surface. Particles begin to saltate when aerodynamic forces are sufficient to entrain a particle into the flow, or impacting particles eject surface particles into the flow. Saltation occurs when \( u_s \geq u_{st} \), where the threshold friction velocity, \( u_{st} \), is a function of the air density \( \rho_a \), particle density \( \rho_p \) and particle size \( d \), as defined in Equation (2.13).

Suspension occurs when the wind speed associated with turbulent eddies \( u_\ast \) is large compared to the particle fall velocity, which for particles in Stokes flow would be given by \( \frac{\rho_p g d^2}{18 \mu} \). Hence, for suspension to occur, the following criteria has to be satisfied:

\[
 u_\ast \geq a \frac{\rho_p g d^2}{18 \mu} \tag{6.1}
\]

When the constant \( a \) is greater than 10, particles behave as flow tracers and are independent of their initial conditions (Bagnold, 1941). For \( 10 > a > 1 \), particles are influenced both by turbulence and their initial conditions, and are in modified saltation (Nalpanis, 1985). When \( a < 1 \), particles are assumed to be in pure saltation (Jensen and Sørensen, 1983). The domain bounded by Equations (2.13) and (6.1) at a temperature of \(-10 \degree C\) and ambient pressure of \(835 \text{ hPa}\) is shown in Figure 6.1.

Experiments in the wind tunnel were carried out over various snow surfaces at a range of \( u_\ast \). The airborne particles therefore varied in size, and as a comparison of the regions in Figure 6.1 and results in Figure 6.2 shows, several experiments with low mean particle sizes and high \( u_\ast \) are likely to have shown more characteristics of suspension than saltation.

As was shown in Figure 2.5, modelling suspension requires a base concentration. This basal condition can only be provided by direct measurements or modelling of the saltation process. However, those experiments where larger particles were observed to be in motion can be simulated directly, as the dominant mode of mass transport should be saltation. Four experiments were therefore chosen as suitable
Figure 6.1: Modes of snow transport as $f(u_*, d)$. Boundaries are drawn at the limits of aerodynamic entrainment with different values of $A$ and at the transition between pure saltation and turbulence-influenced transport. In this sketch, $T_a = -10$ °C, ambient pressure is 835 hPa.

Figure 6.2: Ranges of $u_*$ and $d$ in wind tunnel experiments. $d$ is the mean value from a gamma fit to size data recorded by the SPC, and $u_*$ is the value measured by the miniairs above the saltation layer.

for modelling. These are the experiments of 17/01/2005, 20/01/2005, 02/02/2005 and 21/02/2006. The mass fluxes and number of observed particles in the selected experiments is shown in Figure 5.3.

Further examination of the fluxes measured at each point show that the material detected by the SPC
6.1. DRIFTING SNOW IN THE WIND TUNNEL

Figure 6.3: Mass fluxes in selected wind tunnel experiments. The mean mass flux measured by the SPC is shown as a function of $u_*$ for several experiments. The numbers alongside each point are the total number of particles measured at each point.

includes a significant proportion of small particles. Rates of particle transit through the SPC detector are shown in Figure 6.4. These are the actual rates of detection, not corrected for area.

Figure 6.4: Rates of particle detection at the SPC. Data is shown for selected wind tunnel experiments.

The test on 21/02/2006 was excluded from the threshold data (Chapter 4) because most of the
measurements before drift started showed low $z_0$, suggesting the instruments were above the internal boundary layer. However, once drift started, $z_0$ increased significantly and the boundary layer interface is apparently then above the instruments (see Figure 3.15). As measured particle transport rates for that test were low, it is assumed that the influence of saltating particles on the velocity profile is low, and hence the $u_*$ and SPC mass-flux data can be used.

The frequent passage of small particles through the SPC domain would, at first glance, suggest that these experiments are not as suitable as was thought, based on the mean dimensions. However, the contribution of a particle to the total mass flux is proportional to the number of particles of size $d$, or $d^3$. Doubling the particle diameter increases the mass by 8 times; the contribution of smaller particles to the total mass transport, even if they occur frequently, can be relatively low. As the size distribution is known, equation (6.1) allows the particle motion to be checked, and so the relative mass flux from saltating and suspended particles can be established. The mass flux from saltation in the selected experiments is typically more than 98% of the total mass flux, making these good experiments to simulate using a saltation code.

### 6.2 Modelling saltation

Saltating particles follow a parabolic path through a wind field. The path is determined from a force balance on the particle, using the equations of motion, given as Equations (2.21) and (2.22). A simple time-marching routine can be used with the equations of motion to calculate the progress of the particle from the surface, into the boundary layer and back to the surface;

\[
x(t + \delta t) = x(t) + (\ddot{x}(t) \cdot \delta t) \\
z(t + \delta t) = z(t) + (\ddot{z}(t) \cdot \delta t)
\]  

(6.2)

(6.3)

whereby the location at time $t + \delta t$ (in either $x$ or $z$) is given by the displacement from the previous location in an interval $\delta t$. The trajectory that is followed is therefore strongly influenced by the initial conditions.

The horizontal mass flux through a vertical plane 1 m wide is the product of the number of particles in motion per unit area of surface, their mass and their mean velocity. This gives

\[
\dot{Q}_{\text{salt}} = N_{\text{salt}} \frac{4}{3} \pi \left( \frac{d}{2} \right)^3 \rho_p r
\]

(6.4)

where $N_{\text{salt}}$ is the total number of saltating particles. Depending on the surface conditions, $N_{\text{salt}}$ could include a mixture of aerodynamically entrained and rebounding particles, and so both processes should be included in a model. As was noted by Doorschot and Lehning (2002), the presence of rebounding particles, rather than just entrainment, could significantly increase the mass flux over a surface. $N_{\text{salt}}$ and $\pi$ are both strongly affected by the surface conditions, the form of the boundary layer and modelling assumptions, and these boundary conditions are often described by probability distribution functions. The mass flux for one set of conditions therefore has to be combined with other cases from across the corresponding probability distribution function to obtain the total flux. Particles passing through the boundary layer also extract momentum from the air, and so alter the boundary layer. This in turn alters the entrainment of snow particles, resulting in a feedback process, which has to be modelled (Raupach, 1991).

#### 6.2.1 Drag forces

Although other forces also act on the particles, Gauer, 1999, amongst others) have demonstrated that lift and drag forces are dominant. Once particles have left the surface, experiments by Rice (1991) show a minimal influence of shape on total transport rates, and analysis of particle trajectories by Maeno et al. (1979) showed that snow particle accelerations were consistent with Stokes flow around a spherical particle. The relative velocity $u_r$ is defined as

\[
u_r = \sqrt{(\dot{x} - u(z))^2 + (\dot{z})^2},
\]

(6.5)
where $\dot{x}$ is the particle horizontal velocity, $u(z)$ the wind speed at height $z$ and $\dot{z}$ the particle vertical velocity. Doorschot (2002) calculated the drag coefficient ($C_D$) from the relative Reynolds number, $Re_r = u_r d/\nu$ using a formulation which is valid over a wide range of $Re_r$, and at low Reynolds numbers is the same as that found for Stokes flow;

$$C_D = \frac{24}{Re_r} + \frac{6}{1 + Re_r^{1/2}} + 0.4.$$  \hspace{1cm} (6.6)

6.2.2 Particle initial conditions

The trajectory of a saltating particle is dominated by the initial conditions. These are often defined in terms of the ejection angle $\alpha_{ej}$, and the ejection speed $v_{ej}$, as these can be determined from analysis of video or photographs of the particle trajectories at the surface. A sketch of the trajectories followed by various combinations of particle size and initial conditions is shown in Figure 6.5.

![Figure 6.5: Particle trajectories in a boundary layer. $u_* = 0.5 \text{ m s}^{-1}$, $z_0 = 0.021u_*^2/2g$. Particle diameters are 300 and 500 $\mu$m, $\alpha_{ej} = 25^\circ$ or $45^\circ$, ejection speeds are 2.5 or 3.5 $u_*$; details of combinations used are shown next to the relevant trajectories. No feedback is included between particles and the boundary layer velocity profile.](image)

It is clear from a comparison of the trajectories shown in Figure 6.5 that the correct ejection velocity and ejection angle is required, to obtain the correct vertical distribution of particles. Underestimating the ejection angle, or the ratio $v_{ej}/u_*$, reduces the height that particles reach, and reduces the hop-length of the particles.

Data from literature show a wide range of ejection speeds and ejection angles. These are summarised in Table 6.1. Experiments with fresh or natural snow covers are lacking; most surfaces in literature tended to be described as ‘smooth’, ‘flat’ or ‘carefully prepared’. By comparison, although wind tunnel surfaces were generally smooth, there was some topology on scales larger than individual grains which might change the behaviour of the surface. However, as experiments used generally granular material, the information might be applicable to a snow saltation model.

The data presented in Table 6.1 suggests that for $u_* \leq 1.5u_*$, particle ejection angles are dependent on particle diameters, and show occasional large deviations from the mean. This is particularly the case for the experiments of Araoka and Maeno (1981) with snow, in which a particle seeder upstream may have resulted in a large particle size variation, compared to other experiments with less fragile ice particles, sand or seeds. Consideration of the trajectories shown in Figure 6.5 also suggests that...
Table 6.1: Saltation ejection parameters from literature. Measurements are all from wind tunnel experiments. Data collected over snow surfaces are highlighted. Particle sizes and ejection angles are given as \( \bar{d} \pm \sigma(d) \) and \( \bar{\alpha}_{ej} \pm \sigma(\alpha_{ej}) \); \( d_1 - d_2 \) and \( \alpha_{ej,1} - \alpha_{ej,2} \) indicates a range of values without statistical information. Ratios of \( \frac{v_{ej}}{u_*} \) and \( \frac{u_*}{u_{st}} \) are those given in the literature.

<table>
<thead>
<tr>
<th>Surface</th>
<th>( d ) (( \mu )m)</th>
<th>( u_* ) (m s(^{-1} ))</th>
<th>( \alpha_{ej} ) (°)</th>
<th>( \frac{v_{ej}}{u_*} )</th>
<th>( \frac{u_*}{u_{st}} )</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>flat sand</td>
<td>118,188</td>
<td>34 - 41</td>
<td>4</td>
<td></td>
<td></td>
<td>Nalpanis (1985)</td>
</tr>
<tr>
<td>hard snow surface with seeding</td>
<td>100</td>
<td>65, [10-110]</td>
<td>( \alpha_{ej} )</td>
<td></td>
<td></td>
<td>Araoka and Maeno (1981)</td>
</tr>
<tr>
<td>old snow with seeding</td>
<td>480 ± 240</td>
<td>0.3</td>
<td>25 ± 15</td>
<td>2.9</td>
<td>1.5</td>
<td>Nishimura and Hunt (2000)</td>
</tr>
<tr>
<td>ice particles</td>
<td>2800 ± 300</td>
<td>0.35</td>
<td>18 ± 13</td>
<td>5.4</td>
<td>1.0</td>
<td>Willetts and Rice (1985),</td>
</tr>
<tr>
<td>mustard seeds</td>
<td>1800</td>
<td>0.51</td>
<td>23 ± 18</td>
<td>2.1</td>
<td>1.5</td>
<td>summarised in McEwan et al. (1992)</td>
</tr>
<tr>
<td>sand</td>
<td>[150 - 250]</td>
<td>39 ± 26</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>[250 - 355]</td>
<td>25 ± 15</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>[355 - 600]</td>
<td>19 ± 13</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The ejection parameters must be influenced by particle size. If an ejection angle of 25° were uniformly imposed, particles larger than 150 \( \mu \)m in diameter would rarely reach the height of the SPC or be seen in field measurements, yet they are seen to form a significant portion of the observed particles in these cases. Conversely, if an ejection angle of 45° were assumed, then large particles would become more frequent as the distance from the surface increased. This increase in particle size with height is not seen in nature; measurements over snow surfaces in the field and in wind tunnels show a decrease in the main and maximum particle diameter as height above the surface increases (e.g.; Schmidt, 1982; Nishimura et al., 1998; Nishimura and Nemoto, 2005).

The ejection angles from literature are plotted as a function of the particle diameter in Figure 6.6. The literature suggests that for nominally granular particles at \( u_* < 2u_{st} \), the major influence on particle ejection angle is the particle diameter. A fit is also plotted in Figure 6.6 which takes the form of a decay curve, constrained to a maximum ejection angle of 75° and minimum of 20°. The fit is given by

\[ \alpha_{ej} = 75 - 55(1 - e^{-d/175}) \]  \( \text{(6.7)} \)

where \( d \) is the particle diameter in \( \mu \)m. An uncertainty of ±50% is also plotted, i.e. \( \alpha_{ej} \) is approximately \( 20° \pm 10° \) at \( d > 750 \mu \)m, and \( 40° \pm 20° \) at \( d \approx 175 \mu \)m, which encompasses the majority of observations.

The literature summarised in Table 6.1 also shows that the ejection speed is not constant, but is also not a particularly strong function of the particle size. Instead, the ratio \( \frac{v_{ej}}{u_*} \) seems to decrease as \( u_* / u_{st} \) increases. Results are plotted in Figure 6.7. This data was used to parameterise the ejection speed as a function of \( u_* / u_{st} \), assuming an exponential decay in the ejection speed with increasing \( u_* / u_{st} \). Because the data set is limited, the parameterisation was constrained to \( \frac{v_{ej}}{u_*} = 3.5 \) at \( u_* / u_{st} = 1 \). The resulting fit to the data is given by

\[ \frac{v_{ej}}{u_*} = 3.5 - (1 - e^{-5(\frac{u_*}{u_{st}} - 1)}) \] \( \text{(6.8)} \)

Note also that the coefficients have been limited to 1 decimal point. The standard deviation of this function is also approximately ±0.5\( v_{ej} \), and is plotted in Figure 6.7.

6.2.3 Stochastic processes

Saltation does not occur over surfaces of completely uniform particles. The particles which are entrained are of different size, and can experience different initial conditions, as show in Figure 6.6. Correct simulation of these trajectories and the associated mass transport requires that the range of the
ejection angle, ejection speed and particle diameter be defined using some kind of probability density function (PDF). The mean ejection angle and speed can be modelled using Equations (6.7) and (6.8),

\[
\alpha_{ej} = 75 - 55 \left(1 - e^{-d/175}\right),
\]

where \(d\) is the particle diameter in \(\mu m\). The dashed lines are \(\alpha_{ej} \pm 50\%\).

**Figure 6.6:** Particle ejection angles from literature. Data is plotted for the literature survey given in Table 6.1. The bold line is an approximate fit to data, given by \(\alpha_{ej} = 75 - 55 \left(1 - e^{-d/175}\right)\), where \(d\) is the particle diameter in \(\mu m\). The dashed lines are \(\alpha_{ej} \pm 50\%\).

**Figure 6.7:** Particle ejection velocities from Nishimura and Hunt (2000). Data is summarised in Table 6.1. The bold line is an approximate fit to data, given by 

\[
v_{ej}/u_* = 3.5 - (1 - \exp(-5(u_*/u_t - 1))),
\]

where \(u_*\) is the particle threshold friction velocity. The dashed lines are \(v_{ej} \pm 50\%\).
and a PDF generated by assuming some degree of deviation from this mean to give $P(\alpha_{ej})$ and $P(v_{ej})$. Nishimura and Hunt (2000) suggest a joint-normal distribution, while Doorschot and Lehning (2002) assumed a log-normal distribution for $P(\alpha_{ej})$ and $P(v_{ej})$. The size of the entrained material can also be defined using a log-normal distribution (e.g.; Budd et al., 1966), adjusted to the actual particle size local to the measurements. An example of a generic log-normal particle size distribution, with different numbers of elements ($n(d)$) is shown in Figure 6.8.

![Figure 6.8: Generic particle size distributions. Log-normal probability density functions (PDFs) of particle size diameter. The mean and standard deviation of each distribution is constant; $d = 0.3\text{ mm}$, $\sigma(d) = d/3$, but the number of points within each PDF changes. The distributions are plotted with 1, 3, 5, 7, 9, 13 and 17 points. The number of points is shown at the most frequent diameter for each PDF.](image)

The trajectory is then controlled by 8 variables. These include $u_{*}$; $u_{*t}$; $z_0$; air density; air viscosity; and 3 PDF’s for $d$, $\alpha_{ej}$ and $v_{ej}$. Each element of the PDF has to interact with the other, giving a total number of combinations $n(d) \times n(v_{ej}) \times n(\alpha_{ej})$. Each PDF is defined independently of the other. To simulate field measurements, an extra PDF to describe the variation of $u_{*}$ might also be required, but this is neglected for wind tunnel measurements.

### 6.2.4 Particle-wind interaction

As the particles pass through the air, there is an exchange of momentum with the surrounding flow. Examples of the horizontal and vertical accelerations experienced by two particles of different sizes are shown in Figure 6.9.

Horizontal acceleration in these cases is generally positive once the particle is a few mm above the ground, then starts to reduce before the particle reaches a maximum height and is then negative for a larger proportion of the descent than the ascent. These results agree well with those seen in measurements by Araoka and Maeno (1981). Vertical acceleration is always negative, in which case the initial vertical velocity must dominate the height that the particles reach above ground.

The momentum balance between saltating particles and the wind at a height $z$ above the surface is given by

$$\rho u_{*}^2 = \tau_a(z) + \sum_{i=1}^{n} m_i (u_{ij}(z) - u_{ij}(z)),$$  \hspace{1cm} (6.9)
where \( u_* \) is that observed above the saltating material, \( \tau_a \) is the local shear in the fluid, \( m \) the particle mass and \( i \) the index of a particle crossing through the height \( z \), and ↑ and ↓ indicate ascending and descending particles, respectively (Raupach, 1991). This momentum balance can be solved by tracking particles during their trajectories and locally modifying the wind field. The wind field modification then alters the wind shear in the saltating region, and this alters the number of particles that are entrained (Owen, 1964). This feedback can be captured in a model by resolving the process as it develops with distance, which is analogous to fetch in field measurements (Anderson and Haff, 1991). Alternatively, to avoid the computational load of tracking multiple particles, the system can instead be assumed to be in equilibrium, and results iterated until they converge on a stable wind field and transport rate as in Doorschot and Lehning (2002).

### 6.2.5 Entrainment rates

The rate of particle entrainment from the bed is generally assumed to be proportional to the excess shear stress acting on the surface (\( \tau_a \)), over and above that required to cause drift (\( \tau_* \)). Anderson and Haff (1991) suggested that the number of particles entrained per second in a patch 1 m² could be modelled as

\[
n_a = \zeta (\tau_a - \tau_*) ,
\]

where ‘\( \zeta \) is of order \( 10^5 \) grains Newton\(^{-1} \) s\(^{-1} \) [per square meter of sand]’ (Anderson and Haff, 1991), and \( \zeta \) was assumed to be constant. Taking the surface area of the grain as \( \pi d^2/4 \), and calculating the number of grains over a porous surface of area 1 m², and scaling this by the porosity, it is also possible to parameterise \( \zeta \) as

\[
\zeta = \frac{\rho_s}{\rho_{ice}} \times \left( \frac{1}{\pi d^2/4} \right) .
\]

This corresponds to approximately \( 6 \times 10^5 \) grains N\(^{-1} \) s\(^{-1} \) for 1 mm diameter particles on a 50% porous surface, which may be similar to conditions for sand, and rises to \( 2 \times 10^6 \) for 0.1 mm particles over a 95% porous bed, similar to conditions for fine-grained, new snow. The particle entrainment is calculated over a patch 1 m \times 1 m, and the horizontal mass flux calculated for particles passing through
the downstream end of that patch. The resulting mass fluxes that are calculated are therefore for a path 1 m wide, and hence the total transport rate is measured in kg m$^{-1}$s$^{-1}$.

### 6.2.6 Particle-bed interaction

A particle returning to the surface has the potential to eject particles from the surface, or simply to impact and lodge. The energy balance for a particle impacting a loose surface is defined as the sum of the energy carried by the incoming particle ($E_I$), the energy carried by the same particle afterwards ($E_R$), the energy dissipated in the surface ($E_D$), the total energy of the bonds that were broken $E_B$ and the total energy $E_E$ carried by the ejected particles (‘splash’), number $n_E$;

$$n_E = \frac{E_I - E_R - E_D}{E_B + E_E}. \quad (6.12)$$

As $n_E$ must be an integer, the process is not well defined and most attempts to model this use a stochastic approach to the value of $n_E$ (e.g.; Sugiura and Maeno, 2000). Instead of trying to solve this equation, splash could be neglected as most particle impacts result in a single ejection, which may or may not be the rebounding particle (Sugiura and Maeno, 2000). It could then be assumed that a particle simply loses energy on impact with the surface, and the energy of the rebounding particle can be defined in terms of the surface elasticity, $r$, such that

$$E_R = rE_I. \quad (6.13)$$

and hence the amount of energy lost is $(1 - r)$. For a loose, new snow surface, particles have only a low probability of rebounding from the surface; it is more likely that they simply lodge in the pore space, or that the ice matrix collapses as the particles impact. The elasticity is therefore effectively low. Over denser snow, the possibility increases that a particle will rebound as the porosity is lower. The elasticity is therefore defined as equal to the porosity and calculated from the snow bulk density $\rho_s$ and the density of solid ice, $\rho_{ice}$, where $r = \rho_s/\rho_{ice}$. For new snow, $r$ tends to zero, $E_R$ is similarly low, and so the saltation process is dominated by entrainment. For a hard ice surface, $r$ tends to 1 and the loss of energy would be negligible.

### 6.3 A numerical saltation model

Particle trajectories, the number of entrained particles, the wind field modification and thus the drifting snow mass flux are all interdependent. Therefore a model is required which includes these various feedback mechanisms. A model can either allow saltation to develop in the streamwise direction (Nalpanis, 1985; Anderson and Haff, 1991), or an iterative solution can be applied, giving the conditions corresponding to an equilibrium state (Doorschot and Lehning, 2002). The model of Doorschot and Lehning (2002) was developed for use with snow and so will be applied here. The basic model scheme is shown in Figure 6.10. For more information on the various processes, the reader is referred to Doorschot and Lehning (2002).

The saltation model uses two types of variable to describe the saltation process. One type are given as inputs to the model, and describe a particular situation or experiment. These inputs are listed in Table 6.3 and described in Section 6.3.1 below. The other type describes the snow saltation process as described in Section 6.2 and are hard-coded as parameters, and are listed in Table 6.2.

#### 6.3.1 Model inputs

Model inputs can be grouped into 3 categories. These are;

**Weather.** The weather is monitored during an experiment, and the air temperature and pressure are used to calculate the air density and dynamic viscosity. The wind field is described using $u_*$ and the roughness length. $z_0$ is calculated from $u_*$ using Equation (2.20).

**Surface.** The mean and standard deviation of the diameter of the surface particles which are available to be entrained, or a known size distribution. The threshold friction velocity $u_{*t}$ is also given as an input.
6.3. A NUMERICAL SALTATION MODEL

Experiment data
Particle size PDF
Threshold shear

(a) General process

Particle size, \( \alpha_{ej} \) and \( vej \)

saltation type

strong
(Hop length increases, particle rebound dominates)

weak
(Hop length decreases, aerodynamic entrainment dominates)

get the surface shear with only aerodynamic entrainment, \( \tau_e \)

\( \tau_e < \tau_r ? \)

YES

get the mass flux of rebounding particles

Mass flux for this combination of size, \( \alpha_{ej} \) and \( vej \)

NO

get the rebound threshold, \( \tau_r \)

NO (rebounding particles)

Mass flux for this combination of size, \( \alpha_{ej} \) and \( vej \)

Mass flux of saltating snow

(b) Mass flux calculation process

Figure 6.10: The saltation model. (a) Model structure and (b) mass flux calculation loop in the saltation model of Doorschot and Lehning (2002).

Numerics. The size of the PDFs used to describe the particle size \( d \), ejection angle \( \alpha_{ej} \) and ejection velocity \( vej \).

The values used in the simulations are listed in Table 6.3.

The size of PDF used to describe the particle size and take-off conditions influences the final results. Because the PDFs change as the number of classes increase (see Figure 6.8), this number cannot simply be incremented during a calculation, but has to be chosen in advance. For maximum computational efficiency, the number of classes chosen should be the minimum required to give a converged mass flux. Figure 6.11 shows the sensitivity of the total mass flux to the number of classes.

The influence of \( n(d) \), \( n(\alpha_{ej}) \) and \( n(vej) \) is complex. Increasing the number of classes beyond 1 requires a PDF to be used. The PDFs were defined so that for \( n \geq 3 \), the PDF minimum and maximum values do not change, and the PDFs only become more complete. For \( n \geq 5 \), changes reduce even further as the saltation model converges, and the changes in mass flux then represent the complexity of the saltation scheme and the interplay between the processes shown in Figure 6.10.

Convergence is judged as a change in mass flux of less than 1% between simulations with increasing numbers of classes. The calculated total mass flux is converged when 11 classes are used for each of \( d \), \( \alpha_{ej} \) and \( vej \). 11 classes each were used in all further simulations shown in this chapter, unless otherwise stated.

The largest influence on sediment transport rates is usually assumed to be \( u^* \) (Bagnold, 1941; Greeley and Iversen, 1985; Anderson et al., 1991). An example of the total flux rate as a function of \( u^* \) for a generic situation is shown in Figure 6.12. The total mass flux scales with \((u^* - u_{*T})^3 \) with \( R^2 > 0.99 \), as expected from theory. The total mass flux predicted using the original model of Doorschot and Lehning...
Chapter 6. MODELLING WIND TUNNEL SNOW TRANSPORT

Figure 6.11: Influence of altering \( n(d) \), \( n(\alpha_{ej}) \) or \( n(v_{ej}) \) separately on the total mass flux. ‘all’ is the case where \( n(d) = n(\alpha_{ej}) = n(v_{ej}) \). Results are normalised by the case \( n(d) = n(\alpha_{ej}) = n(v_{ej}) = 1 \). \( T_a = -5 \, ^\circ C \), \( P_a = 835 \, hPa \), \( d = 300 \, \mu m \), \( \sigma(d) = 100 \, \mu m \), \( r = 0.05 \). \( z_0 = 0.021u^2/(2 \times g) \). Other model parameters are given in Table 6.2. (a) \( u_* = 0.4 \, m \, s^{-1} \) (b) \( u_* = 0.6 \, m \, s^{-1} \).

is also shown, and in this particular case the mass flux is about half of that calculated by the new, parameterised model. However, this does not imply a constant offset of this magnitude between the two models, as the effect of the parameterisations described in Table 6.2 would be expected to change as the particle size distribution and \( r \) vary.

Figure 6.12: Total mass flux as a function of \( u_* \). \( n(d) = n(\alpha_{ej}) = n(v_{ej}) = 11 \). Results are shown from the new parameterised model as described here, and the model of Doorschot and Lehning (2002). All inputs are as in Figure 6.11.
When comparing simulations with wind tunnel measurements it is important to understand the sensitivity of the model to the input variables. Figure 6.13 shows the sensitivity of the total mass flux to the input variables at different \( u^* \). The threshold for the mean particle diameters in this case would be 0.28 m s\(^{-1}\). Increasing the grain size when \( u^* = 0.4 \) m s\(^{-1}\) initially causes the total mass flux to increase, but as \( d \) increases, the total mass flux decreases as an increasing proportion of the grain size distribution is too massive to be moved. A similar effect is seen with the standard deviation of the grain size. Increasing \( \sigma(d) \) increases the number of large particles, which at low \( u^* \) reduces the number of particles available to drift.

![Figure 6.13: Relative total mass flux as functions of \( d \), \( \sigma(d) \), \( A \) and \( z_0 \) at different \( u^* \). The three cases are \( u^* = 0.4 \) m s\(^{-1}\) (---), \( u^* = 0.6 \) m s\(^{-1}\) (----) and \( u^* = 0.8 \) m s\(^{-1}\) (-----). Values are normalised by the total mass flux from simulations at \( u^* = 0.4 \) m s\(^{-1}\). Other inputs are as Figure 6.11. \( n(d) = n(\alpha_{ej}) = n(v_{ej}) = 11 \).](image)

Figure 6.12 and 6.13 show that the largest influence on the total mass flux comes from changes to \( u^* \). These simulations are driven using friction velocity measurements made with MiniAir anemometers. The measurement uncertainty in \( u^* \) is ±20\%, and so some inaccuracy is possibly introduced by \( u^* \) when comparing experiments and simulations (see section 3.5). The inputs \( d \), \( \sigma(d) \) and \( z_0 \) are less accurately known, particularly because the true size of particles which are entrained is not clear, but the influence of these is smaller than that of \( u^* \).

### 6.3.2 Model parameters

The model used is tuned to the saltation of snow by the choice of several parameter values. These parameters describe the snow saltation system in terms of the processes given in Section 6.2 and are assumed not to vary between experiments in the same way as the weather or surface conditions. Therefore, the parameters are not given as inputs to the simulation, but taken either as fixed, or well defined, for the snow saltation system. Model parameters are described in Table 6.2 and default values given.

The model of Doorschot and Lehning (2002) was originally tuned for relatively large snow particles blowing over dense surfaces, and was not tested with large ranges of particle sizes, as found in the wind tunnel. The parameterisations described in Table 6.2 should increase the range of application of the model. Doorschot and Lehning report little testing of the model sensitivity to these model parameters, other than the surface elasticity, \( r \). As the experiments which were carried out in the wind tunnel used several different types of snow at various temperatures, the parameters listed in Table 6.2 may have changed between experiments. Accurate comparison of simulation and experiment measurements then
Table 6.2: Saltation model parameters. Values in the column headed ‘DL’ given are those originally used by Doorschot and Lehning (2002), apparently based on Nishimura and Hunt (2000). The final column are the values used in this investigation.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Description</th>
<th>DL</th>
<th>New</th>
</tr>
</thead>
<tbody>
<tr>
<td>( u_\theta )</td>
<td>threshold ( u_\theta )</td>
<td>freely chosen</td>
<td>Eqn. (2.13)</td>
</tr>
<tr>
<td>( z_0 )</td>
<td>roughness length</td>
<td>( d/10 )</td>
<td>Eqn. (2.20)</td>
</tr>
<tr>
<td>( \zeta_{ae} )</td>
<td>entrainment coefficient</td>
<td>( 1/(8\pi d^2) )</td>
<td>Eqn. (6.11)</td>
</tr>
<tr>
<td>( \alpha_{ej} )</td>
<td>mean ejection angle</td>
<td>( 25^\circ )</td>
<td>Eqn. (6.7)</td>
</tr>
<tr>
<td>( \sigma(\alpha_{ej}) )</td>
<td>standard deviation of ( \alpha_{ej} )</td>
<td>( 15^\circ )</td>
<td>Eqn. (6.8)</td>
</tr>
<tr>
<td>( v_{ej}/u_\theta )</td>
<td>ratio of ejection velocity to ( u_\theta )</td>
<td>2.5</td>
<td>Eqn. (6.8)</td>
</tr>
<tr>
<td>( \sigma(v_{ej}/u_\theta) )</td>
<td>standard deviation of ( v_{ej}/u_\theta )</td>
<td>0.75</td>
<td>Eqn. (6.8)</td>
</tr>
<tr>
<td>( r )</td>
<td>surface coefficient of restitution</td>
<td>0.35</td>
<td>( \rho_s/\rho_{ice} )</td>
</tr>
</tbody>
</table>

requires that the influence of these parameters is also known. Figure 6.14 shows the change in total mass flux when parameters are changed from their default values.

![Figure 6.14](image)

Figure 6.14: Relative total mass flux as functions of \( \alpha_{ej} \), \( \sigma(\alpha_{ej}) \), \( v_{ej}/u_\theta \), \( \sigma(v_{ej}/u_\theta) \) and \( r \). The three cases are \( u_\theta = 0.4 \text{ m s}^{-1} \) ( ), \( u_\theta = 0.6 \text{ m s}^{-1} \) ( – – – ) and \( u_\theta = 0.8 \text{ m s}^{-1} \) ( – – – ). Mass fluxes are normalised by the same reference case as Figure 6.13.

The parameter with the strongest influence is the surface elasticity, \( r \). For \( r < 0.4 \), the mass flux variation is small. As the surface elasticity is increased beyond this level, the mass flux increases exponentially. This agrees with the results of Doorschot and Lehning (2002). This is due to the switch from so-called ‘weak’ to ‘strong’ saltation, where hop lengths increase as a particle rebounds from the surface, rather than decrease. In strong saltation, the mass flux from rebounding particles dominates over entrainment.

### 6.3.3 Profiles of drifting snow

In its original form, Doorschot and Lehning’s model delivered the total mass flux of saltating snow. By comparison, measurements during wind tunnel experiments gave only the mass flux at a fixed
height above the surface, and so the form of the flux profile above the ground is required if experiments and simulations are to be compared. To compare the results of simulations and measurements in the field, Doorschot et al. (2004) assumed that the mass flux profile could be generated by scaling the total mass flux by a profile function (Equation 4.3) and integrating this over a small height to gives the mass flux through a detector. However, that approach ignored the fact that the saltation model calculates the particle trajectories, and so the time that a particle takes to pass between two points at different heights can be obtained directly. This approach is described below.

The model calculates the integral mass flux for each combination of grain size $d$, ejection angle $\alpha_{ej}$ and ejection velocity $v_{ej}$. Each of these values has a certain probability of occurrence associated with it; these are denoted $P(d)$, $P(v_{ej})$ and $P(\alpha_{ej})$ respectively. To establish the relative horizontal mass flux at a particular height interval $\Delta z$, extending from $z_1$ to $z_2$, the particle trajectory is analysed to find the time $t$ that an ascending ($\uparrow$) or descending ($\downarrow$) particle stays in a certain height interval, relative to the total time-of-flight, $t_f$, such that

$$P(\Delta z) = \frac{\Delta z}{t_f} \left( \frac{dt}{dz} \right) \uparrow + \frac{dt}{dz} \downarrow .$$

As $\Delta z$ can be large compared to an individual trajectory, this is evaluated for each height in the trajectory that falls within $\Delta z$. The horizontal mass flux for a combination of conditions in a particular height interval is then given by

$$Q(d, \alpha_{ej}, v_{ej}, \Delta z) = Q(d, \alpha_{ej}, v_{ej})P(\Delta z).$$

A mass flux profile can then be built up by first calculating $P(\Delta z)$ for several layers at different heights above the surface. Multiplying this by the total mass flux gives the mass carried in each layer. Dividing the mass carried in each layer by the height of the layer gives the local mass flux. A vertical layer interval of 3 mm was chosen for the mass flux profiles as this is larger than the change in trajectory height from one combination of particle classes to another; this prevents the influence of a single particle class from distorting the profile (see Figure 6.3).

### 6.3.4 Estimating SPC transport rates

The mass flux seen by the SPC can be estimated by interpolation of the mass flux profile to the mid height of the SPC. As with the total mass flux, the mass flux at the SPC is a function of the number of classes. This is shown in Figure 6.15. At $u_s = 0.4$ m s$^{-1}$, no particles reach the height of the SPC if the size of only one PDF is varied. Only if PDFs are created for the particle size, ejection velocity and angle together, are any particles modelled as reaching the height of the detector. The simulations converged after 9-11 classes, the same number of classes required for the total mass flux. At $u_s = 0.6$ m s$^{-1}$ the modelled SPC flux stabilises after 7 classes each. Given that both the total and SPC mass flux should be converged at all $u_s$, 11 classes for each distribution were used in the simulations reported here.

The mass flux through the SPC is also a function of the model inputs. The dominant input for the global transport rate was $u_s$, and as is shown in Figure 6.16, it has a very strong influence on the mass flux modelled at the height of the SPC. Changing $u_s$ by a factor 3.5 changes the flux at the height of the SPC by 5 orders of magnitude, and also increases the percentage of the total flux that is at the height of the SPC from less than 0.01% to about 1% of the total. Figure 6.16 also shows the results of simulations using the model of Doorschot and Lehning (2002), and it can be seen that the variation in the mass flux predicted at the height of the SPC between the two models, is much greater than the variation in total mass flux between the models.

The other model inputs, particularly the grain size distribution, are important when calculating the SPC mass flux. The sensitivity of the SPC flux to those, and $z_0$, is shown in Figure 6.17. The dependency of the simulated mass flux through the SPC on the input values follows the same trends as seen for the total saltating mass flux. The input with the most influence is the particle size PDF, and then the roughness length. The particle size distribution affects the number of particles which can be entrained, and in some situations the number will be zero as $u_s$ is too low to move particles. The simulated mass flux is also particularly sensitive to the particle size when the mean particle size is smaller than 400
Chapter 6. MODELLING WIND TUNNEL SNOW TRANSPORT

Figure 6.15: Influence of changing \( n(d), n(\alpha_{ej}) \) and \( n(v_{ej}) \) separately on the modelled SPC mass flux. ‘all’ refers to the case where \( n(d) = n(\alpha_{ej}) = n(v_{ej}) \). Other inputs are as Figure 6.11. Model parameters are given in Table 6.2.

Figure 6.16: SPC and total mass flux as a function of \( u_* \). \( n(d) = n(\alpha_{ej}) = n(v_{ej}) = 11 \). Results are shown for the new parameterisations and the original mode of Doorschot and Lehning (2002). Inputs are as Figure 6.11.

\[ \mu \text{m}, \text{ showing a rapid decrease as the particle size decreases. This is of particular importance when comparing measurements and simulations, as this size of particles is typically found at the SPC.} \]

The model parameters also influence the mass flux measured at the height of the SPC. The relative mass flux as a function of the parameters is shown in Figure 6.18. In comparison to the results shown in Figure 6.14 for the total mass flux, the flux at the SPC is more sensitive to the model parameters, probably because the gradient of flux with height is high at the the height of the SPC. The surface elasticity is not the most influential parameter at the height of the SPC; instead, the ejection angle and...
6.3. A NUMERICAL SALTATION MODEL

Figure 6.17: Relative SPC mass flux as functions of \( d, \sigma(d), A \) and \( z_0 \) at different \( u_\ast \). The three cases are \( u_\ast = 0.4 \text{ m s}^{-1} (-) \), \( u_\ast = 0.6 \text{ m s}^{-1} (- -) \) and \( u_\ast = 0.8 \text{ m s}^{-1} (- - -) \). In all cases, \( n(d) = n(\alpha_{ej}) = n(\nu_{ej}) = 11 \). Other inputs are as Figure 6.11. Results are normalised by the SPC mass flux from simulations at \( u_\ast = 0.4 \text{ m s}^{-1} \).

The speed are more important. As was shown in Figure 6.5 these are the factors which control the height that particles reach during saltation when saltation is dominated by aerodynamic entrainment.

Figure 6.18: Relative SPC mass flux as functions of \( \alpha_{ej}, \sigma(\alpha_{ej}), \nu_{ej}/u_\ast, \sigma(\nu_{ej}/u_\ast) \) and \( r \). The three cases are \( u_\ast = 0.4 \text{ m s}^{-1} (-) \), \( u_\ast = 0.6 \text{ m s}^{-1} (- -) \) and \( u_\ast = 0.8 \text{ m s}^{-1} (- - -) \). In all cases, \( n(d) = n(\alpha_{ej}) = n(\nu_{ej}) = 11 \). Other inputs are as Figure 6.11. Mass fluxes are normalised by the SPC flux in the same reference case as Figure 6.13.
6.4 Simulations of experiments

The inputs used in simulations of the experiments are listed in Table 6.3. The weather conditions and $u_*$ are obtained directly from the experiment records. The threshold friction velocity $u_{*t}$ was obtained from the mean hydraulic diameter of surface particles using Equation (2.13) with $A=0.18$. The particle sizes, given in the table as a mean value and a standard deviation are actually given as a complete, discrete PDF, based on the photographs taken of the surface particles. The grain PDF is extended by interpolation to include particles of the size detected by the SPC. Size resolution is about 40 $\mu$m for particle diameters less than 1 mm, and about 250 $\mu$m at larger diameters, which introduces a rounding error of about 10% in particle diameter or 30% in particle mass. The particle size PDF had 21 classes for all but the test of 17th January 2005, when the large size range required 25 classes. The ejection angle and ejection speed were both described using PDFs of 11 members.

Table 6.3: Inputs to the saltation model. Data was measured during experiments. The grain size PDF was given as an input; $\bar{d}$ and $\sigma(d)$ are illustrative.

<table>
<thead>
<tr>
<th>Date</th>
<th>$P_a$ (hPa)</th>
<th>$T_a$ ($^\circ$C)</th>
<th>$u_*$ (m s$^{-1}$)</th>
<th>$u_{*t}$ (m s$^{-1}$)</th>
<th>$\bar{d}$ (mm)</th>
<th>$\sigma(d)$ (mm)</th>
<th>$\rho_s$ (kg m$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>17/1/2005</td>
<td>837</td>
<td>-3.5</td>
<td>0.26 - 0.81</td>
<td>0.74</td>
<td>2.17</td>
<td>1.24</td>
<td>301</td>
</tr>
<tr>
<td>20/1/2005</td>
<td>831</td>
<td>-2.2</td>
<td>0.31 - 0.61</td>
<td>0.35</td>
<td>0.58</td>
<td>0.33</td>
<td>162</td>
</tr>
<tr>
<td>02/2/2005</td>
<td>840</td>
<td>-7.2</td>
<td>0.21 - 0.36</td>
<td>0.31</td>
<td>1.43</td>
<td>0.41</td>
<td>88</td>
</tr>
<tr>
<td>21/2/2006</td>
<td>826</td>
<td>0.3</td>
<td>0.31 - 0.50</td>
<td>0.39</td>
<td>1.12</td>
<td>0.48</td>
<td>100</td>
</tr>
</tbody>
</table>

6.4.1 Mass flux profiles

A mass flux profile obtained using the model is plotted in Figure 6.19. This plot shows that when only saltation is considered, mass transport is constrained to the lowest 10 cm of the boundary layer. Also, there is no localised maximum in mass transport at some height away from the ground, as suggested by Pomeroy and Gray (1990), but instead the majority of the mass is transported in the lowest 1 cm of the saltation layer. However, the relative transport shown in figure 6.19(b) does however suggest that at higher $u_*$, the transport some distance from the ground could be larger than that directly next to ground. The transport at the height of the SPC is typically less than 1% of the transport rate in the slice closest to the surface.

6.4.2 Particle size distributions

The trajectory analysis that is used to calculate the mass flux at a particular height can also be used to establish the particle size distribution. This data is also available from the SPC for the wind tunnel measurements. A comparison of the particle size distribution at 50 mm above the surface from the simulations and SPC is shown in Figure 6.20 for the experiment of 20/1/2005. Because the simulation and SPC use different size ranges and so cannot be directly compared, the data are grouped in Figure 6.20(b) into fine, medium, coarse and oversized particle size groups, where oversized particles are those greater than the maximum size resolved by the SPC. Mass flux data can therefore be compared to the ‘capped’ data series shown in Figure 6.21.

Figure 6.20 shows that the measured particle size distributions are skewed towards large numbers of small particles, but the total mass that is carried by these smaller particles is small compared to that of the medium and coarse fractions. Also, although large particles are included in the surface particle size PDF, these particles are not always convected to the height of the SPC. Similar distributions of frequency and mass are seen in both the experiments and simulations. There is no obvious trend towards an increase in particle size with $u_*$, suggesting that the increase in mass transport is due mainly to the increased numbers of particles being entrained from the surface.
Figure 6.19: Simulated mass flux profiles for the test on 20/1/2005. These simulations were driven using data from the experiments given in Table 6.3. The position and height of the SPC detector plane is shown for comparison. (a) Absolute values, and (b) relative values.
Figure 6.20: Measured and simulated particle size distributions from the experiment on 20/1/2005. (a) Raw results from SPC and simulation. Note that the diameter classes are not identical in the experiment and simulations. (b) Data grouped by sediment size. Data in the left-hand column are from the SPC, and those on the right are from simulations. Simulation data is scaled so that 100% of particles are below the maximum size observable by the SPC (780 µm). The upper row shows the relative frequency of observation, and the lower row the relative contribution to the mass flux.
6.4.3 SPC mass fluxes

A comparison between simulated and observed mass fluxes for the 4 selected experiments is shown in Figure 6.21. The flux predicted through the SPC by the simulations is shown, as is the mean and maximum flux recorded by the SPC during the experiment. The mean flux shows the likely equilibrium transport rate while the maximum value demonstrates both the range of observations, and the highest possible transport rate of the surface.

The effect of large particles is also considered; in the data series ‘capped’, all particles larger than 780 µm which reach the height of the SPC are capped to that diameter for purposes of calculating the mass flux, but their true size is used to calculate their trajectories. As was seen in Figure 6.20 these particles can carry a significant amount of mass, and taking this into account reduces the mass flux through the sensor by 30-40%, but only at higher \( u_\ast \). For most of the experiments shown here, this is not an important source of error.

The agreement between simulations and measurements varies considerably. Results for the individual experiments show the following:

17/1/2005 This measurement was made over sieved depth hoar. Simulations at \( u_\ast \approx u_{\ast t} \) do not result in any particles reaching the height of the SPC, but as \( u_\ast \) increases to more than 0.8 m s\(^{-1}\), the saltating particles reach 50 mm above the surface. At this speed, the mean and maximum flux rates observed by the SPC bracket those predicted by the model.

20/1/2005 This was a measurement over a natural snow surface. In the two best cases, mean measured flux rates at the SPC are 50-70% of those predicted, but again well bracketed by the mean and maximum fluxes. Data from the second measurement plateau are lower than the other 3 measurements, and are only about 1% of the predicted rate, but this is possibly influenced by the low number of particles passing through the SPC.

02/2/2005 These results are difficult to interpret, as there is only one data point where the anemometers were below the internal boundary layer interface height (see Figure 5.15). Simulation results underpredict the one measured mass flux value by an order of magnitude, and the predicted peak particle size is 300-400 µm, compared to 150-200 measured by the SPC.

Figure 6.21: Simulated mass transport rates at the height of the SPC for selected tests. Values of input variables are given in Table 6.3.
Simulations results agree well with the measured data, except for that at the lowest \( u_\ast \). This value was actually the result of one, 780 \( \mu \)m diameter particle passing through the detector (see Figure [6.3]), and serves as a useful illustration of the stochastic and random nature of drift.

A further simulation was made of the mass flux that would be observed if the SPC were positioned closer to the surface. Results are shown in Figure [6.22] for the experiment of 20/1/2005 with the virtual SPC at 50, 25 and 10 mm above the surface.

![Figure 6.22: Influence of virtual SPC height on results. (a) Simulated mass transport rates for the experiment of 20/1/2005 at the height of the SPC and at \( z = 25 \) mm. (b) Particle sizes at 25 and 50 mm at \( u_\ast = 0.54 \) m s\(^{-1}\). Values of input variables are given in Table 6.3.](image)

Results from these simulation show that the higher transport rates closer to the surface are associated with smaller particles. While this is contrary to field observations, which show a generally reducing particle size with height (Nishimura and Nemoto, 2005), it is consistent with the physics of the saltation model, where larger particles travel further from the surface, as shown in Figure [6.5]. In field experiments the mean diameter is reduced by the presence of smaller particles in suspension or modified suspension, which are not included in the model used here. While the simulation results would suggest a lower threshold at 10 or 25 mm compared to that measured at 50 mm, the effect of turbulence in transporting smaller particles away from the surface probably reduces this effect in the real world.

### 6.4.4 The role of fetch

An important implication of the simulations is that the observed SPC mass flux appears to agree reasonably well with the expected equilibrium transport. In effect, the limited fetch of the wind tunnel is not producing a completely different amount of mass transport to that expected over an infinitely long, flat surface, with no depletion of the surface. By comparison, Liston and Sturm (1998) expected about 500 meters of uniform fetch would be required to achieve 95% of the final equilibrium fluxes of drifting snow. Other field measurements for soil erosion, reported by Zobeck et al. (2003), show that similar distances might be required for an equilibrium flux of soil. Both Liston and Sturm and Zobeck et al. found an exponential rise with distance \( x \) toward some equilibrium transport rate \( Q_\infty \), such that the transport rate at \( x \) was given by

\[
Q(x) = Q_\infty (1 - e^{-x/\ell})
\]

where \( \ell \) is some characteristic length, which varies according to the height of the measurement and surface conditions. Taking the 500 m suggested by Liston and Sturm (1998), this would imply that after 3m of fetch in the wind tunnel the total snow transport would be less than 1% of the equilibrium rate.
However, Jackson and Cooper (1999) found that on a beach, transport rates were in equilibrium within 11 m, and suggested that ‘fetch distance is unimportant when an adequate sand supply is available’. This may be true for dense sand particles where momentum extraction is high, but for less dense particles (such as ice), some fetch dependence might be expected. In the case of snow drift, given hop lengths of less than half a meter for loose snow surfaces (Kosugi et al., 2004), the extraction of momentum from the boundary layer and convergence on an equilibrium state should occur within a few tens of hop lengths at most, especially if drift is sustained. This gives an equilibrium distance of less than ten meters, especially if the wind tunnel is high enough that the roof does not impede boundary layer growth over the saltating material. This distance is closer to measurements of Arnold (2002), who found that for an unseeded sand surface, measured transport fluxes of 90-355 µm diameter particles, soon after the start of drift (< 1200 seconds) varied from 90% to 120% of the final, long distance equilibrium value after just 3 m of fetch.

6.5 Summary

The snow transport measured in the wind tunnel has been simulated using a snow saltation model (Doorschot and Lehning, 2002). That model was originally applied to cases where transport was dominated by large particles, moving in saltation. The model has been adapted to run with probability density functions describing the surface particle size, ejection angle and ejection velocity. This approach required new parameterisations for the ejection angle and velocity as functions of the particle size, and of the surface elasticity as a function of porosity. This allowed the model to be applied in more varied situations than it was originally tuned for. The model was also adapted to calculate vertically resolved, horizontal mass-flux profiles, data from which could be compared to SPC measurements at 50 mm above the surface.

Results of the simulations show that the measured transport rates at the height of the SPC are very similar to those predicted by the model. The mass flux profiles generated by the simulations show that the transport is concentrated in the lowest 10 mm of the boundary layer. Simulations also transport large particles (> 200 µm diameter) higher than smaller particles, which are not expected to reach the height of the SPC. Capping the effective diameter at the maximum size recognisable by the SPC improves the agreement between measured and observed mass fluxes. These results suggest that the smaller particles that were seen in the wind tunnel are most likely transported by modified saltation, or suspension.
Chapter 7

Summary and Conclusions

If you march your Winter Journeys you will have your reward, so long as all you want is a penguin's egg.

A. Cherry-Garrard, *The worst journey in the world* (1922)

7.1 Summary

Wind tunnel measurements using a dynamic pressure rake showed that the velocity profile over snow can be described using the usual log-law approach. This means that the velocity at a height $z$ can be predicted from knowledge of the surface roughness $z_0$ and the friction velocity $u_*$.

At high enough $u_*$, particles were entrained from the snow surface and were detected using a Snow Particle Counter SPC. The SPC measured the snow particle size and frequency in a small height interval. Time-averaged SPC and $u_*$ data were then used to calculate the drift threshold friction velocity, $u_{str}$, from a regression analysis. This was carried out for more than 15 surfaces over 2 winter measurement campaigns. Measurements were also combined with characterisation of the snow surface using visual characterisation, microscope imaging and penetration resistance.

The drift threshold friction velocity was found to vary from $0.28 \leq u_{str} \leq 0.71 \text{ m s}^{-1}$, and varied more with the particle diameter than with atmospheric conditions such as temperature and humidity. Data was compared to a $u_{str}$ algorithm of Schmidt (1980) based on particle dimensions and geometric information. A version with modified parameters is also used in SNOWPACK (Lehning et al., 2000b). Both of these algorithms agree well with the observed threshold, when a small inter-particle bond is assumed. Surface particle forms vary from dendritic forms to small, rounded particles, and when a hydraulic diameter was used for the particle size, $u_{str}$ calculated using Bagnold's algorithm was about twice that for sand of the same diameter. The threshold parameter $A=0.18$ is increased compared to other granular material by intra-particle cohesion, an effect which is also seen in damp sands. The surface particle sizes correlate well with those seen at the height of the SPC immediately after the start of drift, suggesting that the surface entrainment process is influenced by the surface conditions, and that constant values for entrained particle sizes cannot be assumed.

The roughness lengths of the snow surfaces were also measured. These were found to stay approximately constant before the onset of drift, but vary between each surface from 0.04 to 0.14 mm. This is about an order of magnitude less than seen in field studies, but conversely about an order of magnitude more than would be expected from $z_0$ to particle-size correlations which are typically applied for smooth sand. This result probably comes from the small scale topology which is created during snowfall, which is bigger than grain sizes but still smaller than the roughness which is encountered in field situations, ranging from sastrugi to mountains. Importantly for modelling processes in large, flat areas such as Greenland or high latitudes, there was no notable smooth hydraulic regime. However, the onset of drift alters the surface roughness, causing a similar relationship to that seen for Aeolian drift between roughness length and friction velocity. As drift frequency increased, this converged on $z_0 = 0.021u_{str}^2/(2g)$. This implies that processes in snow drift are similar to those of other materials (Owen, 1964).

In order to investigate the influence of permeability and the microstructure of snow on roughness, measurements were also made over permeable foams. The foam permeability varied from $6 - 160 \times$
10^{-9} \text{ m}^2. By comparison, permeability of new snow is typically in the range 0.1 – 10 \times 10^{-9} \text{ m}^2 (\text{Albert and Schultz, 2002}). It was not possible to measure the permeability of the snow covers that were used in the wind tunnel. Measurements were also made over a smooth, solid wall in order to test the instrumentation performance. Results showed that that the smooth data agreed almost exactly with the universal law of the wall. Foam data showed a variation in \( z_0 \) of more than two orders of magnitude, only by changing the foam permeability. \( z_0 \) of the most permeable foam increased slightly with \( u_\tau \), which is a peculiarity of porous media. The highly permeable foams also had a significant zero-plane displacement, implying a downward transfer of momentum from the boundary layer into the media. The depth of the material was changed from 6 \times \text{ pore size} to 38 times the pore size, which had no clear influence on \( z_0 \) or the displacement height. By comparison, the least permeable foam, which was considered similar to snow, had no notable zero-plane displacement, and the roughness length was stable with \( u_\tau \). These data imply that flat snow is effectively impermeable to shear-driven ventilation, and that flow through the microstructure does not contribute to the surface drag. For this reason, length scales such as the snow specific surface area are not of great use in predicting \( z_0 \). Instead, measurements of the surface topology of snow may be more helpful in predicting \( z_0 \), as the surface length scales and drag are what set \( z_0 \).

Drifting snow in the wind tunnel consisted of particles from 50 to 500 \( \mu \text{m} \) in diameter. The range of \( u_\tau \) was sufficient that the smaller particles, below about 150 \( \mu \text{m} \) in diameter, were probably influenced by turbulence and experiencing suspension. Some experiments were dominated by larger particles, for which the majority of transport was probably by saltation. These experiments were simulated using a Lagrangian saltation model (\text{Doorschot and Lehning, 2002}), which was altered so that the initial conditions of the trajectories were parameterised as functions of the particle size, \( u_\tau \) and \( u_\tau \). This was required as the experiments in the wind tunnel covered a wide range of surface particle size distributions and wind speeds. The model was also rewritten to generate vertically-resolved mass flux profiles, the data from which could then be compared to the SPC measurements which were taken at one height above the surface. Data from the SPC in the selected experiments agreed to \( \pm 100\% \) with the modelled mass transport, which, considering the stochastic nature of drift and the small number of data points available, might be considered good agreement. Mass flux profiles were dominated by transport in the lowest 10 mm or so of the boundary layer, which is what is seen in measurements with sand (\text{e.g. Butterfield, 1991}), but less often in field measurements with snow (\text{e.g. Nishimura and Nemoto, 2005}). The difference is likely due to the effect of suspension to transport small particles away from the surface, leading to reduced vertical gradients.

### 7.2 Discussion

The results and observations outlined in the previous section are useful in a variety of applications. The first of these is for avalanche prediction and warning, and the second is where results can be taken from the wind tunnel and applied directly to terrestrial analogues, which are typically found in high latitudes.

Snow transport by wind is a major contributor to avalanche development and danger. Snow drift and subsequent deposition in lee slopes leads to the formation of wind slab, which can then be released under the right circumstances (\text{Schweizer et al, 2003}). As observer networks cannot cover all regions and slope aspects, and cannot be used for scenario-based modelling, avalanche forecasting is increasingly supported by modelling of the development of the snow cover with time and space, for example with SNOWPACK (\text{Bartelt and Lehning, 2002}) or CROCUS (\text{Durand et al, 2005}). These can then be used to estimate the occurrence of drift, based on local weather conditions. As wind speeds are often only measured at a single height, estimating \( u_\tau \) requires either \( z_0 \) or the relationship between \( u_\tau \) and \( z_0 \). The wind tunnel results show that \( z_0 \) of a flat, snow-covered surface is low compared to that for mountain landscapes, and that the contribution of drift to \( z_0 \) can be neglected. Therefore only a \( z_0 \) representative of the local topography is required to estimate \( u_\tau \). The wind tunnel data does show that the outputs from a snow cover model, namely the grain size and form, can be used to estimate the drift threshold, \( u_\tau \), with reasonable accuracy. Given both \( u_\tau \) and \( u_\tau \), the potential for drift can be checked. Next, the snow saltation model of (\text{Doorschot and Lehning, 2002}) has been modified to work in a range of conditions, and has been shown to deliver accurate predictions of the mass flux at a single height. The model also delivers qualitatively accurate profiles of the dependence of saltation on height.
In high latitudes, fetch conditions approach those used in the wind tunnel, in that the dominant length scale is that of the surface media. Also, because of the dominance of katabatic flows on sloping surfaces, winds tend to be of uniform direction and speed. Literature data for these areas shows a similar $z_0$ to that seen in the wind tunnel, and would therefore be expected to increase with the onset of drift. In contrast to Alpine regions, this could be an important factor when estimating $u_*$ from weather station data. The accurate model behaviour is also useful; where transport distances are longer and particle sizes are smaller, the relative mass transport by suspension increases and has to be considered when calculating mass balance, for example. The basal condition for suspension is saltation, which can be accurately modelled, including particle size distributions. The experiments over foam also partly explain a process found in the deeper snowpacks of Greenland and the high latitudes, ventilation. The wind tunnel data show that in low pressure gradients, the low permeability of the snow surface limits shear-driven throughflow, and where it does occur it is most likely limited to the upper rugosities of the surface. By implication, the air movement found at some depth within the snowpack is driven by pressure gradients, which could be induced by local topology.

7.3 Future work

The work presented in this thesis rests on two assumptions. These are;

Wind tunnel velocity boundary layer. It is presumed that the boundary layer generated in the wind tunnel corresponds to that found in nature. This means that measurements should be applicable to, and comparable with, other, flat-field, long-fetch cases. However, there is relatively little information available about the velocity boundary layer very close to a natural snow surface. In particular, turbulence data is lacking. [Williams et al. (1994)] suggest that the the entrainment threshold for sand is strongly influenced by bursts at the wall, which are driven by turbulence. Wind tunnel flow is unidirectional; [Greeley and Iversen (1985)] and [Paphitis et al. (2001)] demonstrated that for sand, vorticity can reduce the drift threshold velocity; might similar effects have a role in snow drift in complex mountain topography?

SPC performance. It is presumed that the SPC delivers accurate assessments of both the grain size and the mass flux. However, the accuracy on the scale of individual particles is not clear, which could alter the mass flux when only small numbers of particles are in motion.

Several solutions to these problems are apparent. These include;

Field hot-wire anemometer measurements. The form of the velocity boundary layer near to a natural snow surface should be measured using hot-wire anemometry. This would allow velocity measurements to be obtained at high frequencies, giving the mean profile and turbulence spectra. Sonic anemometers are not well suited to this task, having large measurement volumes and significant bulk, which makes it hard to obtain measurements near the surface without influencing flow. Similarly, LDV techniques are not ideal because of their response to both flow tracers and snow particles. Data obtained can then be used to tune the wind tunnel boundary layer.

Wind tunnel hot-wire measurements. $u_*$ and $z_0$ have been calculated from profiles which have been limited to a height range by eye. There is some evidence that the selection is reasonable, particularly from data collected over a smooth wall. This approach is not ideal, because of the difficulty in identifying internal boundary layers or the edge of the logarithmic region. It would be far more convenient to have a direct assessment of $u_*$ from time-resolved, two-dimensional velocity measurements using a hot-wire anemometer (assuming that net lateral flow is zero). This would also allow the value of $\kappa$ to be measured directly, rather than being taken as a fixed value.

Direct particle sampling. The SPC is calibrated to deliver accurate mass flux information, and assumes spherical particles. However, there is currently no information on the size and form of particles that are actually entrained from the surface. The form of particles passing through the SPC is also unknown. This information is required to check the SPC function and the entrained particle size-distribution used in the saltation model. Formvar-coated slides [Schaefer (1941)] could be
placed immediately downstream of the sampling volume of the SPC to collect drifting snow particles, and the captured flakes analysed using image processing techniques. The results could be compared to surface snow photomicrographs, giving a direct comparison of drifting and surface snow.

**Multi-point transport measurements.** The form of the saltation layer could be checked using two (or more) SPCs, mounted either at the same streamwise location at different heights, or the same height and separated along the streamline. This would show if the saltating mass flux changes with fetch, or if the mass flux profile shows the same form as expected. The threshold at a lower height (e.g. 25 mm instead of the current 50 mm) could also be checked, and a more detailed profile would allow more accurate limiting of the boundary layer velocity profiles to those regions with and without drift. However, adding extra instruments into the wind tunnel brings a risk of increased blockage and hence flow distortion, which might influence results. Instead a better option could be to move to an imaging based velocity and particle measurement system, two-phase PIV (see section A.2). This would allow particle tracks to be observed. Saltation and suspension can be clearly distinguished from the form of the particle trajectories, which are parabolic and flow-tracer like, respectively. If snow transport by suspension is significant, modelling the mass transport in the wind tunnel would require the saltation model to be coupled to a suspension model. Two-dimensional turbulence data could also provide shear spectra, allowing a probabilistic approach to the threshold algorithms.

A variety of future wind tunnel experiments are also suggested by this work. These include;

**Streamwise drift development.** It was observed that transport rates are similar to those predicted from an equilibrium saltation code. However, the spatial development of drift in the wind tunnel is unknown and should be measured.

**Immobile surfaces.** Experiments up to now have concentrated on erosion, and ignored phenomena such as crust formation. Long duration experiments at \( u_* < u_{st} \) may lead to the formation of a wind crust on the surface, the development of which could be monitored by the SnowMicroPen. Also, the snow surface could be artificially sintered using a water spray. This would bond the upper surface of the snow and prevent drift, but would alter the surface permeability. A fixed surface would allow the development of \( z_0 \) with \( u_* \gg u_{st} \) to be measured, which would demonstrate that the increase in \( z_0 \) with \( u_* \) over a loose surface is due to drift, and nothing else.

**Surface topology measurements.** Experiments over snow show that the usual particle size to roughness length scaling for smooth granular material does not apply to snow. The aerodynamic roughness of the surface may be better determined by the surface topology, which could be measured on the mm to meter scale.

### 7.4 Conclusions

Of the six questions that this work was designed to investigate, five can be answered categorically.

1. Is \( z_0 \) independent of \( u_* \), until drift starts?
   Yes, as far as is measurable.

2. What length scale in the snow best correlates with \( z_0 \) before drift?
   *Neither the particle size, hydraulic diameter or the specific surface area are good predictors for snow roughness lengths. The snow surface is effectively impermeable, and so surface topology, which was not measured, may be a better indicator of \( z_0 \).*

3. What is the threshold \( u_* \) required for drift, and how does it vary with different snow types?
   \( u_{st} \) varies for each snow surface, and ranges from 0.28 to 0.71 m s\(^{-1}\). \( u_{st} \) can be well predicted using the hydraulic radius of particles and by accounting for inter-particle cohesion. There is no single threshold for a single snow type.
4. When drift occurs, does the boundary layer over drifting snow follow the hypothesis of Owen (1964), so that $z_0 \propto u_*^2$?

Yes. When drift is sustained, the dependency is similar to that for sand and soil. The roughness measured above the drifting material is given by $z_0 = 0.021u_*^2/(2g)$.

5. During drift, does the wind shear in the saltation layer drop to the threshold level, or is it still a significant proportion of the free-stream shear?

Velocities near to the ground change compared to a boundary layer without drift, but because of noisy velocity profiles the shear in the saltation layer could not be quantified.

6. Do we observe a similar mass flux and particle size at a given height to that predicted using the model of Doorschot and Lehning (2002)?

Yes, but this could only be checked at the height of the SPC, and only in experiments where mass transport was by large particles moving in saltation. The assumption of pure saltation in the tunnel is also probably incorrect. Most experiments included particles which were small enough to be in suspension.
Appendix A

Notes on measurement techniques

A.1 Drifting snow detector

A Niigata Electric SPC-S7 was the main drifting snow detector in the wind tunnel. This was the same unit as used by Doorschot (2002), Doorschot and Lehning (2002), Doorschot et al. (2004), Doorschot and Lehning (2002) and Lehning et al. (2002b). The SPC is designed to measure the particle size and thus calculate a mass flux of drifting snow through the detector volume. The technique in general has been shown to be accurate for rounded and abraded Antarctic snow by Nishimura and Nemoto (2005), fresh snow by Sato and Kimura (1991) and Sugihara et al. (1998) and artificial snow by Lehning et al. (2002b). More details of the operating principle are given in Figure A.1 and the development history is described on page 106.

The SPC is calibrated so that a sensor voltage corresponds to a particle diameter. The SPC that was in use in the wind tunnel was originally used with a standard calibration, supplied by the manufacturer. In mid 2006, it was discovered that this calibration differed from the one used by Doorschot (2002), Doorschot and Lehning (2002), Doorschot et al. (2004) and Lehning et al. (2002b). The data was corrected to be comparable to those experiments, and then published in Clifton et al. (2006). In January 2007, the device was recalibrated by the manufacturer, and it was discovered that the sensor gain was lower than expected, causing a drop in voltage for the same particle size, compared to expected. This new calibration data is used throughout this thesis. The various calibration curves for the device are shown in Figure A.1.

The calibration is extended to smaller and larger particle diameters by extrapolation of $log_{10}(E_s)$, where $E_s$ is the sensor voltage, and $log_{10}(d_p)$. The noise level of the sensor is less than 30 mV, and signals lower than this voltage are rejected. This limits the lowest detectable particle to approximately 30 µm.

Sensor voltages were not recorded, only indicated diameters. Data was obtained using the manufacturer’s standard curve. Diameters were converted back to voltages using the curves shown in Figure A.1 and then remapped to the new calibrations as required. Further discussions with the manufacturer revealed that it was possible to record the peak voltage drop of the sensor at high frequencies. As is seen from Figure 3.5, this gives the diameter of each particle as it passes through the sensor plane. Recording the sensor voltage directly in future experiments, rather than the indicated diameter, would reduce uncertainties associated with remapping the data using the curves shown in Figure A.1.

The net effect of these calibrations is limited. The chief impact of the January 2007 calibration, compared to the assumed calibrations, is to increase the particle size. This alters the mass flux as well. Data presented in Clifton et al. (2006) made no significant use of the absolute values of mass flux, with the threshold algorithm being insensitive to changes of several orders of magnitude; the algorithm used is effectively a change-point detection. The values of $u_*$ and $z_0$ changed only slightly, changing some values of $R^2$ or regression lines, but generally not fundamentally altering the results. Finally the conclusions drawn in Clifton et al. (2006) are identical to the conclusions drawn in Section 4.
Figure A.1: SPC calibration curves. Data is the sensor output voltage and particle diameter $d_p$. Three curves are shown; the ‘standard’ SPC calibration supplied by the manufacturer, the curve used by Doorschot (2002), Doorschot et al. (2004), Doorschot and Lehning (2002), Lehning et al. (2002b) and Clifton et al. (2006), and the actual performance as used in this thesis.

A.2 Alternative drifting snow sensors

Several distinct types of sensor have been developed for detecting snow drift and measuring the rates of mass transport. These include:

**Mechanical traps.** Basically porous bags made of fine fabric mesh, these have an opening facing into the wind and physically stop drifting snow. Used since early Antarctic exploration, they integrate drift measurements over time, and must be emptied on a regular basis.

**Aerodynamic traps.** These are typically some kind of nozzle which faces into the wind leading to a settling chamber. In the settling chamber the velocity is low enough that particles drop out of suspension (see Equation 2.7), and the length is sufficient to prevent saltating particles from passing straight through. The most commonly seen types are the ‘rocket’ form of Mellor (1960). These are relatively large, having a diameter of some 10-15 cm and a length of 40-50 cm. The area (and hence the inverse of the mean wind speed) ratio between the nozzle and settling chamber is greater than 1:200. As the settling velocity of snow particles is below 0.3-0.5 m s$^{-1}$ (see Figure 6.1), this allows measurements in wind speeds up to 40 ms$^{-1}$. These have been used since the early 20th Century, for example during the Australasian Antarctic Expedition of 1910-13 (Mawson, 2000).

**Piezoelectric impact sensors.** A drifting particle has mass and velocity and hence kinetic energy associated with it. The impact of a particle on a piezo-electric surface creates a small current, indicating drift. This approach has the potential for high time resolution if high frequency sampling of the output current from the surface is used. Because the device senses the kinetic energy of the impact, knowledge of either the size or velocity of the impacting particles can be used to estimate the other value. Alternatively, the sensor can be connected to an integrator, allowing time averages to be calculated. This has been commercialised as the ‘Sensit’ sensor, for example, as used by Stout and Zobeck (1996) and Wiggs et al. (2004). Another approach is that taken by the ‘Saltiphone’ (Span and van den Abeele, 1991), where saltating grains impact a microphone membrane, although this is apparently insensitive to particles smaller than 200 µm.
A.3. WIND-SPEED MEASUREMENTS

Acoustic impact sensor. As with the piezo-electric sensor, the impact of a particle transfers energy to the detector, but in acoustic sensors the energy is transferred first to a tube wall, which then causes a measurable acoustic signal (Christin et al., 1999). Because the wind also causes a signal, the signal obtained is filtered to remove the low frequency of the wind but retain the higher frequency particle impact noise. This approach has also been commercialised, as the ‘FLOWCAPT’ sensor, and used in field studies by Jaedicke (2001, 2002). Currently the device is only recommended for use as an indicator, delivering qualitative drift measurements (Jaedicke, 2001; Lehning et al., 2002b), and to a maximum wind speed of $\approx 12 \text{ m s}^{-1}$.

Snow Particle Counters (SPC). These are photo-electric devices which count the number of particles passing through a measurement plane or volume, and characterise the particles by size, number and mass. Two different types of SPC have been developed;

Dual-beam SPC. Schmidt (1977, 1984), Gubler (1981) both describe detectors which use the number, size and duration of the interruption of narrow beams of infrared radiation to measure the frequency of snow drift, divide the detected particles into size classes and then estimate the mass flux for the frequency and size of the particles. Although now commonly known as Schmidt-type detectors, these were apparently developed at the University of Washington in the 1960’s (Schmidt, 1977, p. 1). Work in Japan by Sato and Kimura (1991, 1993) led to the development of a dual-beam sensor using a laser diode.

Brown-type SPC. First described in Brown and Pomeroy (1989), this uses Mie scattering and the subsequent reduction of received radiation to count particles passing through a near-infrared beam. A simple statistical model assuming a probability distribution function for the particle size then gives the mass flux. The relative simplicity and low power requirements of this sensor make it ideal for AWS measurements, although the mass flux estimate is only useful if the particle size distribution is known and remains approximately constant. The SPC-S7 is a further development of the Brown-type, which uses the change in illumination of the sensor as a particle transits the measurement plane to estimate the particle size.

Both types of SPC are essentially light gates with added data processing. These light gates are also available from suppliers of equipment for industrial automation, with similar resolution to those used in snow research but at considerable price advantage. Some work should be invested in finding sensors with sufficiently high time resolution to be useful as snow particle counters.

All of these devices deliver a time series of either averaged or instantaneous data, giving at least the presence of drift and then possibly mass flux rates or a qualitative rate (an ‘index’) depending on the device being used. The exception is the mechanical trap, which can only give an integral measurement over time and space.

A.3 Wind-speed measurements

As was noted in section 2.1, $u_*$ is defined either as the time-averaged value of $\sqrt{\overline{u'w'}}$ within the boundary layer, or as part of the fit to a mean velocity profile, as given in Equation (2.1). $u_*$ can then be measured in a wind tunnel using several approaches;

Velocity fluctuation cross-correlation. The two instantaneous velocity components can be measured using several different techniques;

Particle imaging velocimetry (PIV). In this technique, tracers are introduced into the flow and filmed as they move in an illuminated plane. Calculating the displacement in the flow field over successive images at a known frequency allows the flow in the plane to be resolved. This requires good optical access and illumination, which is usually achieved with a laser sheet. Absolute velocity data requires only a known imaging area. Maintaining a laser or other illumination source at low temperatures is difficult because of coolant and high thermal stresses on start-up. However, the technique allows imaging of relatively large regions and can give time-resolved data.
PIV is problematic in two-phase flow (i.e. snow and air) because of the need to differentiate between the tracer and any other suspended or convected phase. Snow itself does not behave as a tracer, as it is often too large to be convected. A solution to this is offered by Dantec Dynamics, consisting of laser illumination and a fluorescent tracer for the air. The velocities of the different phases are obtained by combined PIV and Laser Induced Fluorescence (LIF). Two cameras with different-wavelength filters are used to detect the snow (which simply scatters the laser illumination) and the tracer separately, which fluoresces at a different wavelength. The velocity information can then be obtained for both phases using usual PIV techniques.

**Laser Doppler Anemometry (LDA).** Tracers are also used in this approach. These are used to seed a volume into which two frequency shifted laser beams are focused. The resulting interference patterns are reflected and frequency shifted by the tracers, and the resulting burst as a tracer passes through the sensing volume can be analysed to return particle velocities, requiring no calibration. This is inherently a point measurement. Usually systems are mounted on a traverse outside of the measurement volume and focused into the volume, and so can also be used to generate data over a larger volume. Measuring from outside the volume could also be problematic in a dense, two-phase flow as the laser cannot penetrate the flow to the region of interest, but this can be mitigated by mounting the sensor on a probe within the flow and measuring near to the probe. However, this then requires the probe to be traversed inside the volume, which disturbs the flow. Systems are best described as horrendously expensive.

**Constant temperature hot-wire anemometry (CTA)** This uses the heat loss from a fine wire - of the order of 10 µm diameter - to determine the velocity of flow around the wire. Using a power amplifier across the wire to maintain the resistance of the wire (and thus the temperature, hence constant temperature anemometry) and measuring the power required gives a measure of the power lost by convection, which is a function of the fluid velocity. A calibration is required to give velocity from the measured power loss. A good overview of the technique is given in [Bruun (1995)](#). The $u$ and $w$ velocities can be measured using dual wires mounted orthogonally, an arrangement known as an ‘X-probe’.

**Sonic anemometers** This uses the different time of flight of a sonic pulse along a path because of a superimposed wind velocity to calculate the velocity along that path. Combining data from 3 paths in different planes then allows the 3 orthogonal velocity components to be calculated.

In each case, the mean velocities $\overline{u}$, $\overline{v}$ and $\overline{w}$ can be calculated over a period of time. Removing the time-averaged mean then allows $u'$ and $w'$ to be calculated for each measurement, and $u_*$ can be obtained from $\sqrt{u'^2 w'^2}$.

**Velocity profile fitting.** Alternatively, $u_*$ and $z_0$ can be calculated by a least-squares fit to Equation (2.1). Measurements of velocity profiles in field experiments usually use 10 - $10^3$ seconds of data.

None of the usual measurement systems are ideal. PIV requires considerable work to implement and interpret. LDA is too expensive and difficult to use in two-phase flow. The fine wires used in CTA are fragile and sensitive to particle impacts. Turbulence measurements in particle-laden flow have been made with sonic anemometers [van Boxel et al. (2004)](#). However, sonic instruments tend to be relatively large with path lengths of 5-20 cm, and this raises the chance that they will disturb the flow field in the wind tunnel.
Appendix B

Nomenclature

The following sections detail the symbols used in this thesis. Variables introduced for convenience in derivations and then discarded, are not included. Dimensions are given using the dimensions length \((L)\), mass \((M)\), time \((t)\) and temperature \((T)\).

**Roman symbols**

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Dimension</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A)</td>
<td>(Bagnold’s) drift threshold parameter</td>
<td>(L^2)</td>
</tr>
<tr>
<td>(A)</td>
<td>parameter in Equation (2.16)</td>
<td></td>
</tr>
<tr>
<td>(a)</td>
<td>area of an object</td>
<td>(L^2)</td>
</tr>
<tr>
<td>(a)</td>
<td>parameter in Equation (2.16)</td>
<td></td>
</tr>
<tr>
<td>(a_m)</td>
<td>constant of proportionality in Equation (4.1)</td>
<td>(M t^2 L^{-3})</td>
</tr>
<tr>
<td>(B)</td>
<td>Biot number</td>
<td></td>
</tr>
<tr>
<td>(B)</td>
<td>parameter in Equation (2.16)</td>
<td></td>
</tr>
<tr>
<td>(C)</td>
<td>constant of proportionality in Equation (2.20), (= 2g \frac{d_s}{u^2})</td>
<td></td>
</tr>
<tr>
<td>(C_D)</td>
<td>coefficient of drag</td>
<td>(M L t^{-2})</td>
</tr>
<tr>
<td>(C_L)</td>
<td>coefficient of lift</td>
<td></td>
</tr>
<tr>
<td>(C_M)</td>
<td>coefficient of moment</td>
<td></td>
</tr>
<tr>
<td>(D)</td>
<td>drag force</td>
<td>(M L t^{-2})</td>
</tr>
<tr>
<td>(DN)</td>
<td>mapped dendricity</td>
<td></td>
</tr>
<tr>
<td>(d)</td>
<td>particle characteristic size</td>
<td>(L)</td>
</tr>
<tr>
<td>(d_b)</td>
<td>bond diameter</td>
<td>(L)</td>
</tr>
<tr>
<td>(d_l)</td>
<td>particle mean diameter measured from images</td>
<td>(L)</td>
</tr>
<tr>
<td>(d_h)</td>
<td>particle hydraulic diameter = (4a/p)</td>
<td>(L)</td>
</tr>
<tr>
<td>(d_{SPC})</td>
<td>particle size indicated by SPC</td>
<td>(L)</td>
</tr>
<tr>
<td>(h_s)</td>
<td>saltation height</td>
<td>(L)</td>
</tr>
<tr>
<td>(h_{ss})</td>
<td>snow depth</td>
<td>(L)</td>
</tr>
<tr>
<td>(I)</td>
<td>inter-particle cohesive force</td>
<td>(M L t^{-2})</td>
</tr>
<tr>
<td>(k)</td>
<td>permeability</td>
<td>(L^2)</td>
</tr>
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<td>(k_e)</td>
<td>effective thermal conductivity</td>
<td>(M L t^{-3} T^{-1})</td>
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<td>(L)</td>
<td>lift force</td>
<td>(M L t^{-2})</td>
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<tr>
<td>(\ell)</td>
<td>characteristic size</td>
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</tr>
<tr>
<td>(\ell_{pore})</td>
<td>pore diameter</td>
<td>(L)</td>
</tr>
<tr>
<td>(\mathcal{L})</td>
<td>characteristic length for the saltation system</td>
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<td>(O)</td>
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<tr>
<td>(M)</td>
<td>moment</td>
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<tr>
<td>(N_3)</td>
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<tr>
<td>(p)</td>
<td>perimeter of an object</td>
<td>(L)</td>
</tr>
<tr>
<td>(P_a)</td>
<td>ambient pressure</td>
<td>(M L t^{-2})</td>
</tr>
<tr>
<td>(P_{dyn})</td>
<td>dynamic pressure</td>
<td>(M L t^{-2})</td>
</tr>
<tr>
<td>(Q)</td>
<td>total horizontal mass transport</td>
<td>(M L^{-1} t^{-1})</td>
</tr>
<tr>
<td>(r)</td>
<td>surface elasticity</td>
<td></td>
</tr>
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Re
Re
Re
RH
SP
SSA
T
t
u
v
w
x
y
z
zd
z0
zp
Greek symbols

\[ \alpha \] convective heat transfer coefficient
\[ \sigma_e \] particle ejection angle
\[ \beta \] factor in gamma distribution function (Equation 3.3)
\[ \Gamma \] gamma function (Equation 3.3)
\[ \gamma \] factor in gamma distribution function (Equation 3.3)
\[ \delta_v \] viscosity scale
\[ \lambda \] saltation height parameter
\[ \mu \] dynamic viscosity
\[ \nu \] kinematic viscosity
\[ \Phi \] mass transport vertical profile function
\[ \phi \] porosity (percentage void)
\[ \rho_{air} \] air density
\[ \rho_p \] particle density
\[ \rho_s \] snow bulk density
\[ \sigma() \] standard deviation of a quantity
\[ \sigma_b \] reference shear stress in Equation (2.16)
\[ \tau_a \] aerodynamic shear stress at the surface
\[ \tau_s \] aerodynamic shear stress
\[ \tau_{st} \] threshold shear stress
\[ \tau_r \] rebound threshold shear stress
## Constants

<table>
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<tr>
<th>Symbol</th>
<th>Description</th>
<th>Units</th>
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<tr>
<td>$\kappa$</td>
<td>von Kármán constant, 0.41</td>
<td></td>
</tr>
<tr>
<td>$g$</td>
<td>acceleration due to gravity, $9.81 \text{ m s}^{-1}$</td>
<td>$Lt^{-2}$</td>
</tr>
<tr>
<td>$\pi$</td>
<td>3.14159...</td>
<td></td>
</tr>
<tr>
<td>$\rho_{\text{ice}}$</td>
<td>ice density 917 kg m$^{-3}$</td>
<td>$ML^{-3}$</td>
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</table>
Bibliography


Curriculum vitae

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Employment


Publications


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Davos, 2007