THE IMPACT OF BLOWING SNOW ON SNOW CHEMISTRY

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INTRODUCTION

During and after deposition to open and windswept areas, snow is redistributed by the wind to surfaces that are relatively sheltered from erosive aerodynamic forces. This redistribution impacts the chemistry of snow and snowmelt by transporting snow and its contaminants across the surface and by forming snowpacks whose physical characteristics such as depth, density, particle bonding, crystal habit and porosity vary widely over a basin. The variation of these physical characteristics in windswept conditions has been well documented (for example Steppuhn and Dyck, 1974; Tabler, 1975; Gray, 1978; Langham, 1981; McKay and Gray, 1981; Tabler et al., 1990). Likely in association with this physical variability, a chemical variability has been observed (Davies et al., 1987; 1988; Tranter et al., 1987). Delmas and Jones (1987) measured a reduction of ion concentrations in surface snow after a drifting snow episode on a lake snowcover in eastern Canada, they suggested the nature of

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chemical changes during blowing snow is complex and should be considered when calculating dry deposition to snowpacks.

This paper reviews the processes of snow transport and chemical change during transport, presents recent measurements of changes in snow chemistry during wind transport and compares the measurements to a model of snow chemistry changes caused by sublimation of blowing snow.

**DEFINITION OF THE PROBLEM**

Changes to snow chemistry associated with wind transport of snow may be considered a result of the transport of chemical species between sources and sinks and the transformation of these species along the flow paths. The contaminant transport mechanisms during blowing snow are turbulent transport of chemicals bonded to blowing snow particles, aerosols and as free vapour. Transformations of chemicals in blowing snow are not necessarily conservative with respect to the snow particle. During wind transport snow particles may scavenge or release contaminants as aerosols and vapour and water vapour may sublimate from the particle. The release of contaminants from sources and the absorption by sinks are affected by changes to the snow surface associated with the flux of blowing and falling snow. Using the frame of reference of a control volume of blowing particles above a layer of surface snow, the changes in the concentration of ions in these snows may be found as the result of ion and snow fluxes to the control volume and transformations of ions and ice resident in snow particles within the volume. Figure 1 is such a control volume; aerosols, gases and blowing snow enter from the upwind boundary and exit through the downwind boundary, an exchange of gas and aerosols occurs at the top of the boundary as does wet deposition of ions in snowfall. The bottom boundary is the snow surface where snow and aerosols may be deposited or re-entrained and gases absorbed or released by the surface snow. The system of ion exchange in Figure 1 though complex, is largely controlled by blowing snow transport processes and ion exchange with blowing snow particles.
Neglecting chemical processes in snowfall genesis (considered by Barrie in this volume), the next section reviews blowing snow physical and chemical properties that may impact snow chemistry.

**BLOWING SNOW PROPERTIES THAT IMPACT CHEMISTRY**

**Erosion of Surface Snow**

Erosion of surface snow proceeds when the shear force transferred from the atmosphere to the snow surface exceeds inter-particle cohesive forces and particle inertia sufficiently to support the weight of newly eroded particles (Schmidt, 1980; Schmidt, 1986; Pomeroy and Gray, 1990). This shear force is most effectively transferred by the impact of saltating particles.
Anderson and Haff (1988) modelled erosion of non-uniform, non-cohesive particles showing that the impact of large particles with higher energy trajectories usually "splash" out a number of small particles with lower energy trajectories. This finding may apply to erosion of cohesive snow particles. The resistance to snow particle ejection provided by cohesion increases with the strength and number of snow surface bonds, hence poorly bonded crystals are more likely to undergo ejection from the surface. The measurements of Martinelli and Ozment (1985) suggest a great spatial variability to surface bond strength, often with location on snow dunes and topographic features. The ratio of exposed surface area to particle mass increases dramatically when snow is eroded, as does the ventilation rate of the exposed surfaces.

Blowing Snow Particles

Lamb et al. (1986) demonstrated that contaminant concentrations in falling snow particles are stratified with regard to particle size. Blowing snow particles however, do not usually resemble the original precipitation or snow surface crystals, having been fragmented during ejection and rebound along the snow surface and sublimated when lifted above the surface. Schmidt (1981; 1984) and Pomeroy and Male (1988) noted blowing snow particles are relatively amorphous; a slightly oblong spheroid shape is often observed. Particle density is typically 900 kg/m³ and size decreases with height from a mean radius (of a sphere of equivalent volume) of 100 μm just above the surface to 40 μm at about 2 m height (Budd, 1966; Schmidt, 1982a). The size distribution of blowing-snow particles is skewed towards smaller radii and has been fitted to the two-parameter gamma distribution for suspended (Budd, 1966) and saltating (Schmidt, 1981) blowing snow. The form of the gamma distribution is

$$f(r) = \frac{r^{(\alpha-1)} e^{-\frac{r}{\beta}}}{\beta^\alpha \Gamma(\alpha)}$$
where \( f(r) \) is the relative frequency of a particle of radius \( r \), \( \alpha \) is the distribution shape parameter (dimensionless), \( \beta \) is the scale parameter (m) and \( \Gamma \) denotes a gamma function. The mean snow particle radius is equal to the product of \( \alpha \) and \( \beta \) and the variance of radii is equal to the product of \( \alpha \) and \( \beta^2 \). Whilst a more comprehensive expression using turbulence, snowfall rate and surface condition parameter is desirable, Pomeroy (1988) fitted approximate expressions for the variation of mean particle radius (of a sphere of equivalent volume), \( r_m \) (m), with height, \( z \) (m) and \( \alpha \) parameter with height to measurements Schmidt (1982a) made over a range of heights from 0.05 to 1 m above the surface. The expressions are,

\[
\begin{align*}
  r_m &= 4.6 \times 10^{-5} z^{-0.258} \\
  \alpha &= 4.08 + 12.6 z
\end{align*}
\]

where the coefficient in Eq. 2 has units m\(^{1.258}\) and the coefficient in Eq. 3 has units m\(^{-1}\). Figure 2 shows modelled distributions of blowing snow particle radii for various heights above the surface.

Given a gamma distribution, the mean blowing snow particle mass, \( m_m \), may be found from \( \alpha \) and the mean particle radius as,

\[
m_m = \frac{4}{3} \pi \rho \bar{r}_m^3 (1 + \frac{3}{\alpha} + \frac{2}{\alpha^2})
\]

The difference between the mean blowing snow particle mass and the mass of a blowing snow particle of mean radius is notable, reaching a maximum near the surface where the mean particle mass is 70\% greater than the mass of particle with mean radius.
Figure 2. Two-parameter gamma distributions of blowing snow particle radii typically found at four heights above the snow surface.

Ventilation of, or wind velocity relative to, blowing snow particles is an important parameter to calculation of heat and vapour transfer to the particle and ion scavenging rates. For suspended snow, i.e. above a height of about 0.1 m, the horizontal speed of blowing snow particles is equivalent to the mean horizontal wind speed (Schmidt, 1982a). Assuming blowing snow particles fall with a terminal mean vertical velocity and that inertia restricts their ability to follow the paths of inertia-less fluid points in the turbulent atmosphere, the particle ventilation rate $v_x$ (m/s) may be modelled as consisting of a steady vertical and a fluctuating three-dimensional component. The steady component is found using an expression Pomeroy and Male (1986) developed from Carrier's drag law that relates blowing snow terminal fall velocity to the particle radius. For simplicity, the fluctuating component is found using an empirical approximation of Lee's (1975) calculation of the root mean square turbulent velocity relative to a typical blowing snow particle over a continuous snowcover. Adding the steady
(left side of right hand) and fluctuating (right side of right hand) components of ventilation velocity yields:

\[ V_{r} = 1.1 \times 10^{7} r^{1.8} + 0.01061 u_{z}^{1.36} \]  \hspace{1cm} (5)

where \( u_{z} \) is the mean wind speed (m/s) at height \( z \) and the units of the coefficients are \( \text{m}^{-0.8}\text{s}^{-1} \) and dimensionless respectively. Noting the decrease with height of snow particle radius as shown in Equation 2, the mean component of the ventilation rate decreases with height whilst given a logarithmic mean wind speed profile the turbulent component increases slightly with height and increases with upper-level wind speed.

The particle impact process and sharp wind speed gradients at the snow surface in saltating snow have been hypothesized to generate an electrostatic charge in blowing snow. The phenomenon may interfere with polar radio communication during blizzards and Wishart (1970) reported measurements of a negative electrostatic charge for Antarctic blowing snow particles that are derived from surface snow.

Wind Transport

A major impact of snow redistribution on snow chemistry is the transport of ions carried by snow within and between basins. This transport may occur during or after initial surface deposition. The two primary modes of snow transport are saltation and suspension. Saltation occurs nearest the surface and is the dominant mode at low wind speeds whilst suspension occurs to heights of several tens of metres and dominates at high wind speeds.
-Saltation

Saltation of snow refers to the transport of snow particles bounding along the surface in parabolic trajectories roughly one or two centimetres high and twenty centimetres long. Particles are shattered and abraded during rebound and ejection from the surface, especially if the original crystal is intricate. This shattering may release large aerosols previously contained in snow crystals.

Snow is eroded to and deposited from the saltation layer; these fluxes and the weight of snow supported by the layer are controlled by the balance of atmospheric shear stresses and the snow flux to/from the suspended transport layer above. When the total atmospheric shear stress exceeds that exerted on the non-erodible surface and that required to shatter bonds and expel surface snow particles, saltation may be maintained (Pomeroy and Gray, 1990). The condition at which saltation is just maintained is termed the "threshold" and the shear stress at this point (if all the surface is erodible) is the "threshold shear stress". Shear stress applied to the erodible surface in excess of the threshold level "powers" saltation transport. The balance of shear stresses implied in this argument is related to saltation transport rate \( Q_{salt} \) \((kg/(m\ s))\) as follows,

\[
\frac{Q_{salt} \cdot g}{u_p} = e (\tau - \tau_n - \tau_t)
\]

(6)

where \( g \) is the gravitational constant \((m/s^2)\), \( u_p \) is the mean horizontal velocity of saltating particles - the saltation velocity, \( e \) is the efficiency of saltation, \( \tau \) is the shear stress, \( n \) referring to that exerted on the non-erodible surface and \( t \) to that exerted at the threshold of transport.

The measurements of Pomeroy and Gray (1990) indicated an inverse relationship between efficiency of saltation transport and wind speed; possibly explained by greater kinetic friction resulting from particle impact and ejection at higher wind
speeds. As well, Schmidt (1986) and Pomeroy and Gray (1990) suggested that the saltation velocity is proportional to a wind speed within the saltation layer. Over a complete snowcover, saltation layer wind speed is relatively independent of height and of the wind speed above the saltation layer; evidence for this is the increase in the apparent aerodynamic roughness height with shear stress. The aerodynamic roughness height, \( z_0 \), has been postulated to increase in proportion to the height of saltating particle trajectories, the saltating particles themselves acting as roughness elements (Owen, 1964). Assuming snow particles are ejected at some vertical velocity proportional to the friction velocity, \( u^* \), their maximum trajectory height should be proportional to \( u^* ^2 \). Following this reasoning Tabler (1980), Tabler and Schmidt (1986) and Pomeroy and Gray (1990) have fitted measurements of aerodynamic roughness height and friction velocity during blowing snow to the following equation,

\[
z_0 = c \frac{u^* ^2}{2g}
\]  

(7)

where \( u^* \) is the friction velocity (m/s). Empirical values for the dimensionless coefficient \( c \) are shown in Table 1. Because variable \( z_0 \) during blowing snow, the relationship between friction velocity and upper-level wind speed is not linear, the degree of non-linearity depends upon the value of \( c \). Schmidt (1986) noted that for short time periods, the coefficient \( c \) varies even more widely with snow conditions than is shown by mean values as reported in Table 1.

A saltation transport expression Pomeroy and Gray (1990) derived from the shear stress apportionment theory, observations of wind speed reductions near the saltation layer and measurements of the saltation mass flux is:

\[
Q_{salt} = \frac{0.68 \rho}{u^* g} \left[ u_t^* u^*^2 - u_t^* u_a^* - u_t^* \right]
\]  

(8)
Table 1. Values of blowing snow aerodynamic roughness height coefficient $c$ derived from wind speed profile measurements during blowing snow.

<table>
<thead>
<tr>
<th>Investigator</th>
<th>Conditions</th>
<th>$c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tabler (1980)</td>
<td>70% snowcovered lake</td>
<td>.026</td>
</tr>
<tr>
<td>Tabler and Schmidt (1986)</td>
<td>Snowcovered Wyoming Short-grass Plateau</td>
<td>.063</td>
</tr>
<tr>
<td>Pomeroy and Gray (1990)</td>
<td>Snowcovered Canadian Prairie grainfield</td>
<td>.120</td>
</tr>
<tr>
<td>Pomeroy and others (Unpublished)</td>
<td>Snow and ice covered, boulder-strewn Scottish Highland Plateau</td>
<td>.089</td>
</tr>
</tbody>
</table>

where the coefficient, 0.68 (m/s), reflects the saltation velocity and efficiency, $\rho$ is atmospheric density (kg/m$^3$), and the subscripts $t$ and $n$ to friction velocities refer to the apportionment of shear stress applied to the erodible surface at the threshold and that applied to the non-erodible surface. Typical threshold friction velocities are 0.2 m/s for fresh dry snow and 0.35 m/s for wind-hardened snow (Kind, 1981). The non-erodible friction velocity is usually small for continuous snowcovers. The expression predicts that for a fresh, loose snowcover, saltation transport will begin at low wind speeds and increase only moderately with wind speed, whilst for hardened snowcovers transport begins at higher wind speeds and increases sharply with wind speed. In both cases the relationship between saltation transport and upper-level wind speed is approximately linear.

Implications of this saltation theory are that topography, snow surface structure and wind speed determine whether the threshold condition for saltation is met. The quantity of snow transported in saltation increases approximately linearly with wind speed and the mass concentrations of saltating snow in the atmosphere vary from 0.4 to 0.9 kg/m$^3$. Blowing snow must pass through the saltation layer while being eroded or deposited and saltation shattering may release aerosols from snow crystals, a phenomenon possibly observed by Delmas and Jones (1987).
Suspended snow is that prevented by the upward drag force of turbulent eddies from regularly contacting the surface. These forces are incapable of entraining surface snow (Schmidt, 1982b; Anderson and Hallet, 1986; Pomeroy, 1988) so the source from which turbulent diffusion of suspended snow proceeds is the saltation layer. Hence, without concurrent precipitation, suspended snow can not occur unless the threshold condition for saltation is exceeded. Suspension favours the lifting of smaller particles, leading to a sorting of particle size with height as demonstrated by Eq.s 2 and 3. Snow may be suspended to several tens of metres in height, though the mass concentration of blowing snow in the atmosphere declines sharply with height.

The transport rate of suspended snow, \( Q_{susp} \) (kg/(m s)) is found by integrating mass concentration, \( \eta_z \) (kg/m\(^3\)), over height, \( z \), with reference to the concentration, \( \eta_x \), set to 0.8 kg/m\(^3\) at height \( z_x \), the interface between saltation and suspension, where

\[
Q_{susp} = \frac{U'}{k} \int_{z_x}^{z_b} \eta_z \ln\left(\frac{z}{z_0}\right) \, dz
\]  

and \( k \) is von Kármán's constant (0.4), \( z_0 \) is the aerodynamic roughness height and \( z_b \) is the height of the top of the boundary layer. The vertical gradient of mass concentration is extremely difficult to describe and model for conditions of irregular terrain, incomplete boundary-layer development, intermittent saltation transport, a highly stratified atmosphere and flow around obstacles. For tractability the following discussion assumes a fully-developed atmospheric boundary-layer for which diffusion of snow is in a steady-state, i.e. the momentum transfer variables and the mass concentration of blowing snow at some height are unchanging with time. In practice these conditions are typically found in the lowest metres of the
atmosphere over snow-covered surfaces with several hundred metres of relatively level and uniform upwind fetch.

For these fully-developed conditions, the gradient of mass concentration can be modeled by analogy to one-dimensional laminar diffusion, modified for continuity of mass of sublimating snow particles undergoing gravitational settling in a turbulent atmosphere where:

\[
\frac{\partial \eta}{\partial t} = -\frac{\partial (w \eta - K_s \frac{\partial \eta}{\partial z})}{\partial z} - [\frac{dm_m}{m_s dt}] \eta
\]  

(10)

and \( w \) is the mean vertical snow particle velocity, \( K_s \) is the turbulent diffusivity of blowing snow at height \( z \) and \( \frac{dm_m}{m_s dt} \) is the rate coefficient for sublimation of snow particles, indexed to the snow particle of mean mass. Integrating Eq. 10 for steady-state conditions (\( \frac{d\eta}{dt} = 0 \)) provides an equation similar in form to that developed by Shiotani and Arai (1953) and confirmed by Budd et al. (1966), except that in the following development, \( w^* \) is height dependent,

\[
\eta_I = \eta_1 \left( \frac{z_f}{z_i} \right)^{w^*}
\]  

(11)

The dimensionless diffusion parameter, \( w^* \) is a ratio between the average gravitational settling and upward turbulent velocity of blowing snow particles that is difficult to calculate because of the uncertainty in the degree of "slip" of blowing snow particles against inertia-less fluid points in the atmosphere and the possibility of "turbulent burst" mechanisms for upward diffusion. However, measurements in fully developed blowing snow (Pomeroy, 1988) provide an empirical solution for \( w^* \) as a function of height,
where the coefficient has units of \( m^{1.544} \) s\(^{-1}\). This relationship suggests that as a result of selective diffusion of particles, atmospheric turbulence and particle fall velocity reach a steady balance that is independent of wind speed and snow surface conditions.

The reference height for turbulent diffusion of snow from a mass concentration of 0.8 kg/m\(^3\) (estimated for the saltation layer) increases linearly with friction velocity (Pomeroy, 1989)

\[
z_r = 0.05628 u^*
\]

where the coefficient has units of seconds. The resulting mass concentration of suspended snow decreases logarithmically with height, the rate of decrease diminishing with wind speed. For instance, at heights of 0.1 and 1.0 m the mass concentrations of suspended snow given a 6 m/s 10-m wind speed are 0.04 and less than 0.001 g/m\(^3\) respectively, for a 20 m/s wind speed the concentrations are 200 and 2 g/m\(^3\).

-Mass Flux-

Using numerical techniques based on the preceding development (Pomeroy, 1988) to model saltation and suspension transport, the vertical profile of horizontal mass flux to a height of 10 metres may be calculated for various wind speeds (Fig 3). Mass flux in the saltation layer declines with wind speed and that in the suspended layer increases rapidly with wind speed; hence the height at which the maximum mass flux occurs increases from near the surface at threshold wind speeds to approximately 0.1 to 0.15 m at higher wind speeds (15 to 20 m/s). As a result, the suspended transport rate increases rapidly with wind speed and becomes the dominant mode of transport for wind
Figure 3. Vertical profiles of blowing snow mass flux modelled for fully-developed flow over a continuous snowcover. The wind speed \( U \) is that at a 10-m height. The excessively sharp changes in gradients are a modelling artifact.

speeds above the near-threshold condition. In example, for near-threshold wind speeds saltation comprises almost all transport but at \( u_{10} = 8 \) m/s suspension comprises 75% of the total and more than 90% at \( u_{10} \) greater than about 17 m/s. Substituting the 10-m wind speed, \( u_{10} \) for friction velocity and integrating the mass flux to a height of 5-m provides the total blowing snow transport rate \( Q \) (kg/(m s)) as a function of \( u_{10} \). An approximation of this relationship, valid over the range of \( u_{10} \) from 6.5 (the threshold condition) to 25 m/s, is

\[
Q = \frac{u_{10}^{4.04}}{458800}
\]
with the coefficient having units kg$^{-1}$ m$^{5.04}$ s$^{3.04}$. Note that this approximation is derived only for completely snowcovered and unobstructed upwind fetches of length greater than 500 m. Under such conditions strong winds, even of short duration, can transport vast quantities of snow.

Sublimation of Blowing Snow

Sublimation of ice from blowing snow particles can increase the concentration of contaminants if contaminants are not removed during the sublimation process. Delmas and Jones (1987) suggested that sublimation of blowing snow may release significant numbers of contaminant aerosols from snow particles. However, noting that sublimation of blowing snow is a gradual mass transfer that proceeds from an intermediary quasi-liquid layer surrounding the particle and that thermo and diffusio-phoretic forces during sublimation favour aerosol scavenging rather than release, the more likely tendency is for aerosols to remain incorporated in a blowing snow particle during sublimation. Gaseous transfer from the quasi-liquid particle surface could be significant, depending on the relative concentrations in the atmosphere and in the snow particle.

The dispersed particulate nature of blowing snow means that sublimation cannot be considered as from a "simple" snow surface but from a dispersion of non-uniform particles in a turbulent atmosphere. This discussion therefore starts at the single spherical particle scale and may be integrated up to the snow-storm scale using parameters noted in previous sections.

Sublimation of ice from a blowing snow particle is governed by the laws of conservation of mass and energy. The theoretical equation for evaporation of a sphere in turbulent air is

$$\left( \frac{dm_s}{dt} \right) = 2 \pi D r_m Sh (\rho - \rho_m) \quad (15)$$


where \( r_m \) = radius of the snow particle possessing mean mass, \( m_m \); 
\( D \) = diffusivity of water vapour in the atmosphere; \( S_h \) = Sherwood number - a turbulent mass transfer parameter, \( \rho \) = water vapour density in the atmosphere, \( \rho_m \) = water vapour density at the particle surface. Convective heat transfer to a sublimating spherical particle is expressed as

\[
L_s \left( \frac{dm}{dt} \right) = 2 \pi \lambda_T r_m Nu (T_m - T)
\]

(16)

where, \( L_s \) = latent heat of sublimation; \( \lambda_T \) = thermal conductivity of the atmosphere; \( Nu \) = Nusselt number - a turbulent heat transfer parameter; \( T \) = ambient atmospheric temperature; \( T_m \) = atmospheric temperature at the particle surface. Using the Clausius-Clapeyron equation for the water vapour density difference, Thorpe and Mason (1966) combined the above mass and energy balances to provide a sublimation rate expression for ice spheres. Schmidt (1972) added the term \( Q_r \), equal to the radiant energy received by the particle, to the right-hand-side of Eq. 16. By assuming that sublimating blowing snow particles are in thermo-dynamic equilibrium, Schmidt expressed the mass and energy balance for a particle of mass \( m_m \) in the following form,

\[
\frac{dm}{dt} = \frac{2 \pi r_m \sigma - \frac{Q_r}{\lambda_T T Nu} \left[ \frac{L_s M}{R T} \right] - 1}{\frac{L_s}{\lambda_T T Nu} \left[ \frac{L_s M}{R T} - 1 \right] + \frac{1}{D \rho_s S_h}}
\]

(17)

where \( \sigma \) = ambient atmospheric undersaturation of water vapour with respect to ice; \( M \) = molecular weight of water; \( R \) = universal gas constant; \( \rho_s \) = saturation density of water vapour at \( T \). Lee (1975) confirmed Schmidt's model, with modifications for turbulence; Tabler (1975) and Pomeroy et al. (1990a) confirmed application of the sublimation calculation using field
measurements of the mass balance of wind-blown snow and operational procedures for calculating sublimation. The procedure for determining the environmental variables in Eq. 17 is involved and is fully described by Pomeroy (1988).

Enhancement of ion concentration due to sublimation of blowing snow may be calculated using Eq. 17 in the following framework. The normalized concentration \(C_x(t)\) of an ion \((x)\) in a mass of snow where the mixture changes over time may be expressed as:

\[
C_x(t) = \frac{\left[ \frac{m_x(t)}{m_m(t)} \right]}{\left[ \frac{m_x(i)}{m_m(i)} \right]}
\]

where \(m_x\) is the mass of ion, \(m_m\) is the mean mass of a blowing snow particle, \(i\) represents the initial condition and \(t\) is the time elapsed. If \(m_x\) is constant over time, the blowing snow sublimation rate provides the change in \(m_m\) and hence \(C_x(t)\).

The concentration \(C_x(t)\) is plotted against time in Fig. 4 for conditions of 70% relative humidity, -5 °C temperature, 10 m/s wind speed and 120 J/(m^2 s) incoming solar radiation and a variety of initial snow particle radii. The curve for a radius of 40 μm is representative of the concentration enhancement of blowing snow at heights near 2-m, whilst that for the radius of 100 μm is representative of the concentration enhancement of saltating snow under an identical ambient atmosphere. Figure 4 suggests that a vertical gradient in ion concentration enhancement will develop, a result of the gradient in snow particle size. To show the sensitivity of concentration enhancement to ambient atmospheric conditions, \(C_x(t)\) for a transport time of 30 sec. is plotted against air temperature for three different relative humidities in Fig. 5. There is a sharp decline in normalized concentration as relative humidity increases and temperature decreases. This suggests a gradient of increasing concentration with height will develop because
Figure 4. Concentration enhancement due to blowing snow sublimation as a function of time. Sublimation is calculated for \( u = 10 \, \text{m/s}, T = -5 \, \text{°C}, \, \text{RH} = 70\% \) and \( Q_r = 120 \, \text{J/(m}^2 \, \text{s)} \).

Relative humidity is usually higher and wind speed and temperature lower near the surface during blowing snow. Vertical gradients of particle size, wind speed, air temperature and humidity should cause a notable increase in sublimation rate with height and hence enhancement of contaminant concentrations should increase with the height of the blowing snow particle. For a column of blowing snow, a dramatic enhancement of contaminant concentration may occur as blowing snow sublimes in relatively "mild" and "arid" winter conditions.
The accumulation/depletion of surface snow is the lower boundary of a mass-balance of blowing and falling snow in the atmospheric boundary-layer (Pomeroy, 1989). As shown in Fig. 6, the mass-balance may be applied to a control volume of blowing snow: a column of the atmosphere extending 5-metres in height above a unit area of surface. In this control volume the surface accumulation is the sum of the snowfall and the difference between the snow blowing into and out of the volume, less that snow lost to sublimation within the volume, where:
Control Volume for Blowing Snow Transport

\[ \text{Accum} = Q_{\text{in}} - Q_{\text{out}} - \int_0^t \frac{dN}{dt} \, dz + S \]  

and Accum is the rate of surface snow mass accumulation per unit area, \( Q \) is the horizontal transport rate of blowing snow into and out of the control volume, \( dN/dt \) is the sublimation rate for a unit volume of atmosphere and \( S \) is the downward flux of falling snow.

The measurements of Takeuchi (1980) and considerations of atmospheric boundary-layer development theory (Hunt and Weber, 1979) suggest that the boundary-layer condition where on average the blowing snow entering a 5-m high volume equals the blowing...
snow leaving it, requires a level, snowcovered, uniform upwind fetch of greater than 400-500 m. These fetches occur on northern plains, upland plateaus and ice sheets, hence for many locations in these regions the average snow accumulation in excess of vegetation/roughness element storage is a balance between the quantity of snowfall and losses to blowing snow sublimation. In more irregular terrain and for incomplete snowcovers, the balance of blowing snow entering and leaving a control volume exerts the primary control on snow accumulation. For irregular terrain situations the transport rate algorithms presented must be modified for incomplete snowcover and short fetches. Suitable modifications have not yet been specified, though Tabler and Schmidt (1986) presented methods for distributing the snow drift accumulation behind snowfences and in topographic depressions.

Noting that ions are carried by or situated in the snow considered in the mass-balance, it is clear that a simple wet deposition velocity describes accumulation inadequately and a control volume mass-balance may be constructed for "wet" ion transport. Such a mass balance is,

\[ \text{Accum}_x = Q_{xin} - Q_{xout} - \int_{0}^{5} \frac{dM_x}{dt} \, dz + S_x \]  

(20)

where for ion x carried by snow, Accum\(_x\) is the rate of surface accumulation of ion x, Q\(_x\) is the horizontal transport rate of ion x into and out of the control volume, the third term represents the rate of scavenging of the ion x by snow and S\(_x\) is the vertical flux of ion carried by snowfall into the control volume. Because of strong turbulence, it may be assumed that dry deposition by aerosols will be small compared to the rates of wet deposition during snow redistribution. If the concentrations of ions in snow, blowing snow transport and sublimation rates, snowfall rates and ion scavenging rates are known then the accumulation of ions at the snow surface may be solved for. It is important to consider that whilst in a steady-state scenario
sublimation does not affect the ion transport rate, over a blowing snow event or season where Accum is controlled by topographic or vegetation capacity, the concentration of ions in accumulating snow and hence sublimation must be considered to find the long-term Accum, the "snow-ion holding" capacity of a catchment.

Snow deposited by wind transport has different properties from surface snow that has not blown. Saltation impact, shattering and sublimation of ice reduces blowing snow particle size from the original crystal. In-situ surface snow is subjected to the shattering action of particle impact and in-pack sublimation due to relatively high ventilation associated with the high winds during blowing snow (Langham, 1981). Wind-hardening and densification result. Bonds between deposited snow particles form rapidly from the time of contact. Jellinek (1957) reported that the work to disaggregate deposited particles doubles within one day and triples within three days. Fresh snow deposited with light winds has a typical density of 100 kg/m$^3$. However snow deposited by drifting has densities typically three times this value. As snow is buried by overlying drifts its density increases by compaction from the applied pressure. Tabler (1985) provided an expression based on extensive measurements, for the density, $\rho_d$ (kg/m$^3$) of dry, drifted snow:

$$\rho_d = 522 - \frac{304}{1.485y} (1 - e^{-1.48y})$$

(21)

where $y$ is the depth of snow (m) and the constant 522 has units kg m$^{-3}$. The maximum density which can be achieved is just over 500 kg/m$^3$. Narita's (1971) observations of snowpack structure showed a decrease in the total surface area of snow particles per unit mass; given a density of 100 kg/m$^3$ the total surface area is 25.4 m$^2$/kg, whilst for a density of 300 kg/m$^3$ the area is 21.7 m$^2$/kg. Perla (1985) found a more dramatic change in total surface area with density, suggesting about 45 m$^2$/kg for a
density of 100 kg/m³ and concurring with Narita's (1971) observations for the 300 kg/m³ value.

scavenging During Transport

Snow particles may be blown for distances of 3 to 5 km, resulting in atmospheric residence times of up to 10 min. (Tabler, et al., 1990). In the lowest metres of the atmosphere during blowing snow, water vapour density is normally undersaturated because of strong turbulent mixing. Hence, scavenging of ions by blowing snow is not accomplished by riming and most resembles the scavenging processes of below-cloud precipitation; i.e. it is dominated by vapour transfer. Below-cloud scavenging may significantly contaminate snow particles. For instance, Zinder et al. (1988) suggested up to 80% of contaminants in snow falling by a mountainside were scavenged below the cloud in humid but sub-saturated conditions. Beard et al. (1983) listed several mechanisms of aerosol scavenging by evaporating drops of 60-90 μm radius, that also apply to sublimating blowing snow particles. They are:

1) inertia
2) interception
3) gravitation
4) thermophoresis
5) diffusiophoresis
6) Brownian motion
7) electrostatic attraction/repulsion.

Diffusion of soluble gases such as HNO₃, H₂O₂, HCl, H₂SO₄ and NH₃ to the thin quasi-liquid layers on the surface of snow particles may be added to this list (Dawson and Brimblecombe, 1983; Huebert et al., 1983).

Scavenging by Brownian motion dominates for aerosols of radii less than 0.01 μm; inertial impact scavenging is added to other modes for radii greater than 1 or 2 μm. Na⁺, often carried by larger aerosols, may be subject to inertial impact and scavenged more rapidly than the other ions carried by aerosols (Zinder et al., 1988). SO₄²⁻ is carried by aerosols of 0.1 to 1
\( \mu m \) radius; without condensation, aerosols of this size are scavenged primarily by particle interception, phoretic effects and electrostatic attraction (Hegg and Hobbs, 1983). The ventilation velocity (Eq. 5) may be multiplied by the snow particle projected area to calculate the volume rate of atmospheric interception for a blowing snow particle falling through a turbulent atmosphere containing aerosols. Sophisticated convective diffusion models of aerosol transport to columnar ice crystals such as that of Miller and Wang (1989) can be combined with turbulent mass and energy transfer models for blowing snow particles (Eq. s 15 and 16). Sublimation creates gradients of temperature and water vapour pressure very near the surface of blowing snow particles that may favour thermophoretic and diffusiophoretic scavenging of aerosols. For evaporating, falling snow particles, convective diffusion models suggest collection efficiencies for aerosols from 0.1 to 1.0 \( \mu m \) radius may increase an order of magnitude with a relative humidity decrease from 95\% to 50\% (Miller and Wang, 1989). Murakami et al. (1985) and Sauter and Wang (1989) note that collection efficiency decreases with increasing size and compactness of a snow crystal. The field measurements of Lamb et al. (1986) provide strong evidence that small snow particles are much more efficient than large particles in phoretic capture of aerosols. The collection efficiency of blowing snow particles due to phoresis should increase with height above the surface in response to declining particle size.

Electrostatic charge may increase the aerosol collection efficiency of snow crystals by two orders of magnitude (Wang and Sauter, 1985). When particles are highly charged, electrostatic attraction of aerosols can be an important scavenging mechanism (Sauter and Wang, 1989). The measurements of Murakami and Magono (1983) suggest positively charged snow particles have twice the collection efficiency of negatively charged snow. They note an inverse relationship between the increase in collection efficiency due to electrostatic effects and the snow particle ventilation velocity. Strong electrostatic charges in snow blowing near to the snow surface might result in relatively high aerosol collection efficiencies.
Zinder et al. (1988) measured increases of 50% and 150% in snow particle concentrations of NO₃⁻ and NH₄⁺ respectively as snow fell 1200 m in humid conditions along a mountain slope. These increases are much greater than for ions carried primarily by aerosols. Hence, they suggested that gaseous scavenging operates much faster than aerosol scavenging to falling snow particles. Huebert et al. (1983) noted the scavenging coefficient for HNO₃ vapour is five times larger than that for NO₃⁻ in aerosols, suggesting gaseous transfer is the dominant mode for this ion. SO₂ gas may also be scavenged and oxidised to SO₄²⁻ from highly polluted atmospheres (Hegg and Hobbs, 1983). Dawson and Brimblecombe (1983) suggest the scavenging of gases should be partitioned by their solubility. Given the scavenging coefficient of 25 x 10⁻⁵ for HNO₃ vapour (Huebert et al., 1983), after 30 sec. of falling snow the atmospheric concentration is reduced by less than 1%, though after 10 min. a 16% reduction occurs.

These results should be applied to blowing snow with caution. It would seem for blowing snow that scavenging of gaseous NO₃⁻ and NH₄⁺ would proceed the fastest, followed by SO₄²⁻ in polluted atmospheres and by aerosol Na⁺. It is possible that variations in blowing snow particle electrostatic charge could strongly favour the scavenging of other charged aerosols. Scavenging of aerosols by smaller snow particles is favoured, except when electrostatic charge works to enhance scavenging by larger snow particles near the surface. The blowing snow particle surface area/total mass ratio increases less than proportionately with decreasing mean radius. If gaseous scavenging is proportional to surface area then smaller snow particles will not have notably greater scavenging rates per unit mass than larger particles. Typical scavenging collection efficiencies are between 10⁻² and 10⁻³ for blowing snow sized snow crystals; therefore, except in nearly-saturated water vapour or highly charged electrostatic conditions, snow particle ion concentration changes due to aerosol scavenging should be slower than those due to sublimation.
Observations

An examination of the chemistry of blowing snow, eroding snow and accumulating snow and associated meteorological conditions during wind transport was conducted during March, 1989 in the Cairngorms, Scotland, the results of which are summarised here. The high plateaux of the Cairngorms are particularly suited for such a study as they are subject to frequent and severe blowing snow storms and the precipitation is often heavily contaminated by pollutants from industrialised Europe.

Vertical profiles of wind speed, temperature, humidity, blowing snow particle flux and blowing snow chemistry were monitored in the bottom two metres of the atmosphere over a plateau above the corrie of Ciste Mhearad. Electrostatic charge was measured from blowing snow particles at 0.25 m height, collected in a Faraday cage. When blowing snow was sampled for chemistry, surface snow was also sampled at sites of erosion and deposition. Erosion and deposition sites were distinguished by snow surface hardness, the presence of a drift-crust and dune features. The concentration of various anions and cations was measured from filtered samples using a Dionex ion chromatograph and a Pye Unicam atomic absorption spectrophotometer. A more complete description of the study site, experimental methods, chemical analysis and instrumentation is given by Pomeroy et al. (1990b).

The measurements discussed here were collected under two distinct conditions; Condition A (22 March) developed from snow blowing over a continuous snowcover during snowfall, and Condition B (29 March) developed from snow blowing out of the snow-filled corrie of Ciste Mhearad under clear skies, and sublimating whilst travelling over a locally discontinuous snowcover (ice-covered fell-field exposed between snowpatches). Environmental measurements during the two blowing snow conditions are summarized in Table 2. Selected environmental parameters during the two sets of measurements are graphed in Fig. 7.
Figure 7. Examples of atmospheric parameters recorded during blowing snow chemistry measurements.
Table 2. Summary of environmental observations for blowing snow conditions.

<table>
<thead>
<tr>
<th>Blowing Snow Condition</th>
<th>A</th>
<th>B</th>
</tr>
</thead>
<tbody>
<tr>
<td>Concurrent Snowfall</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Local Snowcover</td>
<td>Continuous</td>
<td>Discontinuous (95% ice/rock).</td>
</tr>
<tr>
<td>Source of Blowing Snow</td>
<td>Cairn Gorm, local &amp; falling snow.</td>
<td>Ciste Mhearad, 300 m distant.</td>
</tr>
<tr>
<td>Age of Snow Surface</td>
<td>&lt; 1 day</td>
<td>1-3 Days</td>
</tr>
<tr>
<td>Electrostatic Charge of Blowing Snow</td>
<td>Negative</td>
<td>Positive</td>
</tr>
<tr>
<td>Wind Speed at 2-m</td>
<td>8-14 m/s</td>
<td>18-24 m/s</td>
</tr>
<tr>
<td>Air Temperature</td>
<td>-5.5 °C</td>
<td>-3.0 °C</td>
</tr>
<tr>
<td>Relative Humidity</td>
<td>85%</td>
<td>89%</td>
</tr>
</tbody>
</table>

Friction velocity (Fig. 7a) and surface aerodynamic roughness height (Fig. 7b) are calculated from vertical profiles of mean wind speed (0.15 to 1.9 m height), fitting the logarithmic case quite well. Relative humidity (Fig. 7d) and air temperature (Fig. 7c) were measured at heights of approximately 0.3 and 0.9 m; height-averaged values are shown. Blowing snow mass flux is shown from two of the particle detectors mounted at heights in the lower suspended layer (Fig. 7e) and well away from the surface (Fig. 7f). Despite higher wind speeds in B, the snow transport rate appears higher in A, this difference is largely due to the limited snow supply in B. Interestingly, during snowfall (A) changes in the mass flux well away from the surface (Fig. 7f) correspond to changes in wind speed (Fig. 7a), however near the surface (Fig. 7e) the mass flux is relative unresponsive to wind speed changes. Concentrations of Cl⁻ (Fig. 7g) and NO₃⁻ (Fig. 7h) in blowing snow are shown from measurements at a height of approximately 0.1 m. Changes in NO₃⁻ concentration during the
measurement period do not appear clearly associated with the changes in other environmental parameters. However, Cl\textsuperscript{-} concentration appear inversely correlated with blowing snow mass flux for Condition B (without snowfall) and positively correlated with mass flux for Condition A (with snowfall). The two peaks in Cl\textsuperscript{-} concentration during snowfall are associated with peaks in upper-level blowing snow flux and wind speed. There are several possible explanations for the behaviour of Cl\textsuperscript{-}. The peaks during "A" were associated with very heavy atmospheric snow loads and relatively strong turbulence; an increase in sublimation rate and resulting greater concentration of contaminants or the transport of more aerosols or more highly contaminated snow to the site are therefore possible explanations for the peaks in Cl\textsuperscript{-} concentration. During "B" wind speeds are high and snow supply is limited. If lulls in mass flux during high wind speeds are associated with restriction of snow supply, then during such lulls blowing snow particles will have, on average, undergone greater travel and sublimation than under conditions of less restricted snow supply. The inverse correlation between Cl\textsuperscript{-} and mass flux during Condition B is a possible manifestation of this effect.

Changes in Blowing Snow Chemistry

Mean ion concentrations in different snow-types (surface snow at sites of erosion and of deposition, and blowing snow at the four heights above the surface) are shown in Fig. 8(a, b). The concentrations are averages of results from approximately 8 distinct samplings (runs in Fig. 7) of snow chemistry; each sampling was conducted over from 10 to 60 min. The values shown are normalized with respect to eroding surface snow values sampled on the plateau. Measurements in Fig. 8 are distinguished between the two blowing snow conditions, concentrations in Fig. 8a refer to the blowing snow with snowfall over a continuous snowcover (condition "A"), and concentrations in Fig. 8b to blowing snow without snowfall over a discontinuous snowcover (condition "B").
8a) Mean of 9 samplings during blowing and falling snow over a continuous snowcover. Blowing snow heights are 0.05, 0.15, 0.46 and 0.86 m.

8b) Mean of 7 samplings during blowing snow over a discontinuous snowcover. Blowing snow heights are 0.12, 0.25, 0.55 and 0.91 m.

Figure 8. Mean ion concentrations in snow during drifting, normalised to eroding snow values. For each ion the sequence of snow-classes from left to right: Eroding surface snow, Blowing snow (4 heights), Accumulating surface snow.
For both conditions A and B, mean ion concentrations are greater in blowing snow than in eroding surface snow except in A for Cl\textsuperscript{-}, NO\textsubscript{3}\textsuperscript{-}, SO\textsubscript{4}\textsuperscript{2-}, Na\textsuperscript{+} and Mg\textsuperscript{2+} at the lowest blowing snow height and for Na\textsuperscript{+} and Mg\textsuperscript{2+} at the second lowest height.

Blowing snow ion concentrations from all four heights were combined and compared to eroding snow values using the Student's "t" test. The probability (0 - 1) that the measured concentrations reflect actual differences between eroding and blowing snow is shown for each ion and both blowing snow conditions in Table 3. The dissimilarities between the two conditions are indicated by the probabilities. For condition B there is approximately 0.99 or better probability that the ions in blowing snow are more concentrated than those in eroding snow. However, for condition A, only NH\textsubscript{4}\textsuperscript{+} and K\textsuperscript{+} display this probability, the next highest being Cl\textsuperscript{-} at 0.84.

Based on an analysis of variance F-test, the probability of a difference between measured concentrations of eroding and accumulating snow is shown for each ion and both conditions in Table 4. For condition A only NH\textsubscript{4}\textsuperscript{+} has a probability greater than 0.8 of being more concentrated in accumulating snows, whilst Mg\textsuperscript{+}, Na\textsuperscript{+} and Cl\textsuperscript{-} have probabilities greater than 0.7 of being more concentrated in eroding snows. For condition B the probabilities for differences between eroding and accumulating snow are greater than 0.9 for Na\textsuperscript{+}, Cl\textsuperscript{-} and Mg\textsuperscript{+}, greater than 0.8 for K\textsuperscript{+} and SO\textsubscript{4}\textsuperscript{2-} and greater than 0.75 for NH\textsubscript{4}\textsuperscript{+}. NO\textsubscript{3}\textsuperscript{-} never shows notably different levels between the accumulating and eroding snows and NH\textsubscript{4}\textsuperscript{+} displays less striking but similar tendencies. Interestingly, in condition A, NO\textsubscript{3}\textsuperscript{-} and NH\textsubscript{4}\textsuperscript{+} concentrations display the greatest enhancement in accumulating snow and in condition B this enhancement is not observable.

To evaluate the significance of the change in ion concentration with height, an analysis of variance was conducted for each ion within conditions A and B with the 4 measurement levels as "treatments". In condition A the ions generally show an increase in mean concentrations with height. The increased ionic concentrations with height are significant at the 99% level of confidence for Cl\textsuperscript{-}, at the 90% level of confidence for Na\textsuperscript{+}, at the 80% level of confidence for Mg\textsuperscript{+}, NO\textsubscript{3}\textsuperscript{-} and NH\textsubscript{4}\textsuperscript{+} and at only the
Table 3. Probability of a difference in ion concentration between eroding surface and blowing snows.

<table>
<thead>
<tr>
<th>Blowing Snow Condition</th>
<th>Cl⁻</th>
<th>NO₃⁻</th>
<th>SO₄²⁻</th>
<th>NH₄⁺</th>
<th>Na⁺</th>
<th>K⁺</th>
<th>Mg²⁺</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>.836</td>
<td>.713</td>
<td></td>
<td>.991</td>
<td>.032</td>
<td>.997</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>.999</td>
<td>.989</td>
<td>.999</td>
<td>.996</td>
<td>.999</td>
<td>.999</td>
<td>.999</td>
</tr>
</tbody>
</table>

All significant differences are for blowing snow concentrations greater than eroding surface snow concentration.

65% level of confidence for K⁺. In condition B, only the anions show a significant increase in concentration with height. Noting that the cations are available for only the two lowest heights; Na⁺ and K⁺ concentrations decrease with height, the decrease in K⁺ being significant at the 75% level of confidence. Increased concentration with height is significant at the 99% level of confidence for SO₄²⁻, NO₃⁻ and Cl⁻ and at the 95% level of confidence for NH₄⁺. The limited range of sampling heights for the cations precludes conclusions regarding their change with height.

Processes of Change in Blowing Snow Chemistry

-Falling & Blowing Snow

Certain processes are indicated by the changes to snowpack chemistry during blowing snow with snowfall over a continuous snowcover (condition A). It is important to note that with heavy snowfall the sequence of snow particle change is not simply eroding to blowing to accumulating snow, but often falling to blowing to accumulating snow. The "dilution" of snow blown over long fetches results in blowing snow being less sublimated than in condition B and accumulating snow not necessarily being derived from eroding snow. The sequence of ionic change with snow type displayed in Fig 8a may be divided into three groups:
Table 4. Probability of a difference in ion concentration between eroding and accumulating surface snows.

<table>
<thead>
<tr>
<th>Blowing Snow Condition</th>
<th>Cl⁻</th>
<th>NO₃⁻</th>
<th>SO₄²⁻</th>
<th>NH₄⁺</th>
<th>Na⁺</th>
<th>K⁺</th>
<th>Mg²⁺</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>.728*</td>
<td>.661</td>
<td>.363*</td>
<td>.977</td>
<td>.758*</td>
<td>.305*</td>
<td>.777*</td>
</tr>
<tr>
<td>B</td>
<td>.934</td>
<td>.130</td>
<td>.807</td>
<td>.784</td>
<td>.943</td>
<td>.856</td>
<td>.892</td>
</tr>
</tbody>
</table>

* Denotes accumulating concentration less than eroding concentration.

1) ions scavenged as aerosols, the sea salt/cloud condensation nuclei ions, Cl⁻, Na⁺, K⁺ and Mg²⁺,
2) ions scavenged gaseously, NH₄⁺ and NO₃⁻,
3) SO₄²⁻, scavenged in both gaseous and aerosol forms.

Accumulating snow shows no enhancement of ion concentrations compared to eroding snow, except for the ions subject to gaseous scavenging, NH₄⁺ and NO₃⁻. SO₄²⁻ shows no difference and is intermediary in this respect. There is an increase in all ion concentrations with height, however, concentrations of sea salt ions in blowing snow near to the surface are lower (except K⁺) than are eroding snow concentrations. Apart from NH₄⁺, the most dramatic increases in ion concentration with height are displayed by the aerosol ions.

A combination of processes is suggested to account for the observed ion concentration profiles. The increase in concentration with height is primarily due to sublimation of ice from snow particles. The rate of sublimation increases with height as a result of the gradients of water vapour deficit, ventilation rate, temperature and ratio of particle surface area to mass. However, some scavenging of NH₃ vapour by blowing snow particles may be indicated by the notably increased concentration with height displayed by NH₄⁺. Saltation particle shattering may release and resuspend relatively large aerosols within the snow crystal. In falling snow these aerosols contain the sea-salt ions that dominate the cloud condensation nuclei. Evidence of this resuspension is manifest in Fig. 5a as reduced concentrations of sea-salts near the surface in blowing snow and
in accumulating snow. This does not necessarily suggest that saltation shattering of eroding surface snow particles releases similar proportions of sea-salt ions. Eroding surface snow is usually somewhat densified and rounded as it is subject to saltation abrasion and wind action before erosion. Though eroding snow is further abraded during transport, it is probably not subject to the severe shattering that a fresh snow crystal would experience upon entering the saltation layer, and hence may retain more of its aerosols. Ions contained in solution within the snow particle such as NO$_3^-$, NH$_4^+$ and SO$_4^{2-}$ would not be dispersed so readily by particle shattering, though, because of the somewhat aerosol nature of SO$_4^{2-}$ it may be intermediary in this respect.

-Blowing Snow over a Discontinuous Snowcover

For condition B, with no snowfall, the opportunity exists for substantial sublimation and ion scavenging because of the long transport distance, restricted snow particle replacement and high wind speeds. The transformation of snow here is less complex than in condition A, being only from eroding snow to blowing snow to accumulating snow. The contribution of ions from the exposed fell-field surface is thought to be small as an ice-layer covered much of the plateau. All ion concentrations in blowing snow are enhanced from eroding snow concentrations and with the exception of NH$_4^+$ and NO$_3^-$ are enhanced in accumulating snow from eroding snow concentrations (Fig 8b). Where fully measured, blowing snow ion concentrations increase with height.

The similarity between concentrations of NH$_4^+$ and NO$_3^-$ in eroding and accumulating snow may be due to vapour transfer of these ions after snow accumulation, as they return to pre-transport levels to balance their atmospheric concentrations. The slower rate of snow transport and accumulation in condition B promotes snowpack exposure whilst rapid accumulations in condition A may have not. Blowing snow eroded from older surface snow should be subject to less saltation shattering than falling snow. In concurrence with this there is no evidence for
resuspension of aerosols in the concentrations of sea-salts at the lowest blowing snow heights; they all exceed the eroding snow concentrations by a large degree. A positive electrostatic charge on this day may have increased the scavenging efficiency for aerosols, contributing to high concentrations of aerosol scavenged ions at lower blowing snow heights.

-Modelling Enhancement of Concentrations: Sublimation

To examine the role of sublimation of ice in changing concentrations of ions in blowing snow, an algorithm has been devised that estimates blowing snow sublimation (Eq. 17) by calculating the sublimation that a column of initially fully-developed blowing snow would experience travelling over a fetch of uniform conditions (Pomeroy et al., 1990b). The initial mean particle mass is "sublimated" for 0.5 sec., reset for the resulting radius, sublimated, reset, etc. Normalized ion concentration at any time, t, is found using Eq.s 17 and 18 with \( m_p = \text{particle mass} \) and \( m_x(i) = m_x(t) \). Time, t, is the fetch distance divided by particle speed, assumed to be equal to wind speed at the chemistry sampling height. Initial conditions are shown in Table 2 for condition B. The wind speed and initial particle mass depend on height and are set to values predicted for the sampling heights. It is assumed that similar conditions may have extended for 1000 m back to the base of Ciste Mhearad. Table 5 shows initial wind speed and particle mass and the resultant \( C_x(t) \) at times representing fetch distances of 300 m and 1000 m. The two fetches represent an "envelope" of the concentration effects due to sublimation during transport from the possible surface snow sources to the measurement site.

The envelope of normalized concentration due to sublimation and the normalized concentrations of measured ions are graphed against height above the snow surface in Fig. 9. The normalized concentrations of \( \text{NO}_3^- \), \( \text{SO}_4^{2-} \), and \( \text{NH}_4^+ \) all fit within the envelope, however the other ions display values that exceed the envelope, particularly at the lowest height.
Table 5. Normalized concentration of ion, $C_x(t)$, due to sublimation of ice from blowing snow particles: $T = -2.6 \, ^\circ C$, RH = 89%.

<table>
<thead>
<tr>
<th>Height m</th>
<th>$U_z$ m/s</th>
<th>$r_m$ µm</th>
<th>$C_x(t)(300 , m)$</th>
<th>$C_x(t)(1000 , m)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.12</td>
<td>9.8</td>
<td>98</td>
<td>1.15</td>
<td>1.66</td>
</tr>
<tr>
<td>0.25</td>
<td>12.6</td>
<td>80</td>
<td>1.17</td>
<td>1.82</td>
</tr>
<tr>
<td>0.55</td>
<td>15.7</td>
<td>60</td>
<td>1.25</td>
<td>2.45</td>
</tr>
<tr>
<td>0.91</td>
<td>17.6</td>
<td>50</td>
<td>1.33</td>
<td>3.48</td>
</tr>
</tbody>
</table>

The sensitivity of the sublimation rate to variation in atmospheric parameters over the fetch should be considered in assessing Fig 9. For instance, the sublimation response to a step change in relative humidity is skewed in that an increase of 5% RH reduces ion concentrations by 25% while a decrease of 5% RH raises concentrations by 41%. However, the vertical profiles of concentration for the exceeding ions suggest a process other than sublimation is responsible for at least part of their values at low heights. Furthermore their values are not likely to be sampling artifacts as they are reflected in high accumulating snow concentrations for these ions. It is possible that blowing snow particles near the surface are scavenging aerosols, and as $Na^+$ displays the greatest excess concentration, scavenging large aerosols. A mechanism that would favour this scavenging to snow, blowing just above the surface, is electrostatic charge which being positive should be relatively effective (Murakami and Magono, 1983), and if generated in the saltation layer is strongest near to the snow surface. Noting this supplementary role of scavenging processes, the concentration enhancement due to sublimation is the right order of magnitude to account for the higher concentrations of "conservative" blowing snow ions such as sulphate and the nitrogen compounds and, accounts for a large part of the enhancement of other ions.
Figure 9. Normalised mean concentrations of ions in eroding surface and blowing snow (Condition B) and normalised concentrations modelled using blowing snow sublimation rates for two possible upwind fetches: 300 and 1000 m.

**CONCLUSIONS**

Physical processes during blowing snow may have profound impacts on snowpack chemistry, by not only transporting contaminants across the landscape but by altering their concentrations in snow. Impact and shattering of snow particles during erosion from the surface exposes and may release aerosols imbedded in the crystals; these aerosols are easily resuspended by the turbulence generated during blowing snow. An electrostatic charge develops on snow moving near the surface and strengthens the forces favouring aerosol scavenging by blowing snow particles. Atmospheric turbulence preferentially lifts smaller particles and hence sorts blowing snow particle size by height; this size gradient can lead to vertical gradients of sublimation and ion scavenging. Wind transport of snow within and between basins redistributes chemicals deposited by both wet and dry deposition over the winter season, and strongly affects
the winter ionic input to small basins in windswept areas. Sublimation of ice from blowing snow proceeds at rates sufficient to enhance chemical concentrations in snow particles several fold over typical transport distances in open landscapes. The sensitivity of sublimation rates to humidity and air temperature means that snow transport distances and ion concentration enhancement are strongly dependent on the concurrent meteorological conditions. After undergoing transport and sublimation, wind-blown snow accumulates in drifts whose location and form are dependant on topographic exposure, vegetation, upwind snowcover supply, wind speed and direction.

Wind transport provides opportunities for further scavenging of atmospheric chemicals by snow. As blowing snow rarely occurs in a saturated atmosphere, scavenging proceeds at rates that are slow compared with scavenging by snow in clouds. Theoretical evidence and below-cloud scavenging studies suggest that scavenging by blowing snow is dominated by gaseous transfer, though phoretic forces due to sublimation and attraction by electrostatic charge may promote notable aerosol scavenging by blowing snow particles in certain favourable situations.

Measurements in Scotland indicate striking changes in snow chemistry during wind transport. Ion concentrations in blowing snow are found to increase with height, and at a height of 1-m can be three times the concentrations in eroding surface snow. This concentration enhancement is sometimes reflected in accumulating snow, where ion concentrations can be twice that in eroding snow. However, during blowing snow with snowfall some ions have concentrations near the surface in blowing snow at and in accumulating snow that are diminished in comparison to those in eroding snow. The degree and nature of ion concentration change in wind-blown snow varies primarily with the atmospheric form of the ion, presence of snowfall, structural change in the snow crystal and opportunity for snow particle sublimation.

During snowfall, the relatively low concentrations of sea-salt ions in blowing snow near the surface may indicate release of aerosols from falling snow as relatively intricate falling snow crystals are shattered upon entering the saltating layer of blowing snow. Concomitant low concentrations of these ions in
accumulating snow suggests that sea-salt aerosols are resuspended after release from saltating snow and that falling snow particles can undergo a significant degree of chemical change in the saltation layer before deposition to the snow surface. Converse to the aerosol loss there is evidence for gaseous scavenging of NH$_4^+$ by blowing and falling snow, evidence supported by enhanced concentrations in accumulating snow.

Without snowfall, rather than release of aerosols, there is evidence for scavenging of sea-salt aerosols by blowing snow near the surface. A strong, positive electrostatic charge held by near-surface blowing snow particles contributes to this aerosol scavenging and the added ion concentration enhancement. Without contribution from falling snow, enhanced concentrations of ions in blowing snow are usually reflected in accumulating surface snow, however, the ions subject to gaseous transfer, NO$_3^-$ and NH$_4^+$, return to eroding snow concentrations after accumulation. This chemical change upon deposition may be due to a gaseous flux of these ions from the new snow surface to the atmosphere, in response to atmospheric concentrations and volatilization in the snowpack.

Enhanced ion concentrations in blowing snow and the increase in ion concentration with height are largely due to sublimation of blowing snow particles. Predictions from a model of blowing snow sublimation that account for vertical gradients of snow particle size and wind speed in calculating the enhancement of contaminant concentrations match measurements of NO$_4^-$, SO$_4^{2-}$ and NH$_4^+$ in blowing snow (without falling snow) quite well. Similar concentration gradients develop during blowing snow with falling snow, though addition of falling snow with its distinctive chemistry modifies the gradient somewhat. Sea-salt ions can be subject to aerosol scavenging and release during blowing snow and NO$_3^-$ and NH$_4^+$ are subject to relatively rapid gaseous scavenging and volatilization, hence, amongst the ions measured in blowing snow SO$_4^{2-}$ is the most "conservative" in it behaviour, displaying and conserving the enhancement of concentration due to blowing snow sublimation most faithfully.
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