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Transport and sublimation of snow in wind-scoured alpine terrain

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ABSTRACT Measurements of blowing snow transport over wind-scoured terrain in the Scottish Highlands indicate mass fluxes several orders of magnitude less than for transport over complete snow covers. A model of snow transport processes indicates that vertical transport of blowing snow out of the near-surface atmospheric boundary-layer and sublimation of blowing snow within this boundary-layer deplete the supply of snow, with important implications for downwind accumulations of snow in forested areas and topographic depressions. Aspects of blowing snow in alpine areas such as wind speed profiles and aerodynamic surface roughness are also addressed.

INTRODUCTION

The redistribution of snow by wind in alpine areas is of great importance to avalanche generation, recreational pursuits, micro-scale water supply and climate and the temporal and spatial distribution of basin runoff. Dyunin & Kotlyakov (1980) noted the effect of sublimation on snow transport in the mountain environment, suggesting that sublimation would limit the seasonal mass flux at heights above the saltation layer. Studies of blowing snow flow over alpine ridge crests in Davos, Switzerland have helped to describe the jets of snow at the ridge crest (Föhn, 1980; Schmidt et al., 1984; Meister, 1989) with regard to modelling the accumulation of cornices in snowy mountain basins (Meister, 1987; 1989). Other studies have concentrated on the distribution of atmospheric precipitation and its deposition to mountainous terrain (Tesche, 1988). Recent interest concerns the accumulation of snow at the forest-edge and in topographical depressions and the chemical load of this wind-blown snow (Pomeroy et al., 1991).

The study reported here examines conditions where the snow cover becomes wind-scoured during blowing snow; such conditions develop in open areas of high wind exposure or low precipitation. Blowing snow flow over wind-scoured snow is not well described by transport equations developed for flow over complete snow covers without corrections for sublimation and vertical transport loss as recognized by Dyunin and Kotlyakov (1980). Simple corrections can permit the evaluation of snow redistribution and sublimation necessary to understand winter hydrology in the alpine zone.

BLOWING SNOW THEORY

Pomeroy et al. (1991) and Tabler et al. (1990) list an expression for the blowing snow transport rate as a function of wind speed. The expression was developed by modelling saltation and suspension of snow to a height of 5 metres over an extensive, level and completely snow covered surface (Pomeroy, 1989; Pomeroy and Gray, 1990) and is:

$$Q_{0\text{-}5} = \frac{u_{10}^{4.04}}{458800} \tag{1}$$

where $Q_{0\text{-}5}$ is the transport rate of blowing snow through a unit width extending to 5 m height (kg m^{-1} s^{-1}) and u_{10} is the mean wind speed at 10-m height (m s^{-1}). The expression and the model it derives from have been successfully used to calculate blowing snow transport to snow fences on the Alaskan Arctic Coastal Plain (Tabler et al., 1990; Tabler, 1991) and to hedgerows on the Canadian Prairies (Pomeroy and Gray, 1991). Both environments are broad plains with simple terrain geometries and continuous snow covers during the modelling periods. In these situations the atmospheric boundary layer near the surface may be considered fully-developed and the flux of blowing snow to any differential volume within the boundary layer approximates a steady-state.

Saltation transport of snow occurs in a layer several centimetres above the surface, where snow particles bound along the surface in curved trajectories. The saltation transport rate is a result of the partitioning of atmospheric shear stress into that required to erode snow particles from the surface (t), that applied to non-erodible surface elements (n) and that available to eject particles from the surface. When saltation flux is in balance with atmospheric shear stress, Pomeroy and Gray (1990) found the following expression valid for the Canadian Prairies:

$$Q_{salt} = \frac{(0.68\ m\,s^{-1})\ \rho\ u_t^*}{u^*\ g}\ [u^{*2} - u_n^{*2} - u_t^{*2}] \tag{2}$$

where Q_{salt} is the saltation transport rate (kg m^{-1} s^{-1}), ρ is the atmospheric density (kg m^{-3}), $u*$ is the friction velocity (m s^{-1}) and the subscripts t and n refer to the portions at the transport threshold and applied to the non-erodible surface elements respectively. The measurements of Takeuchi (1980) suggest that appropriate conditions for equation (2) are found after flow has developed over 100-200 m of uniform snow covered fetch.

The friction velocity controls the wind speed profile in blowing snow, affecting both the intercept and slope of the profile. Several studies (Schmidt, 1982; Pomeroy and Male, 1987) found mechanical turbulence to dominate over buoyancy effects, hence the log-linear model is used. However the aerodynamic roughness height, z_0, is not fixed during blowing snow but is proportional to the trajectory height of saltating particles, postulated by Owen (1964) to be proportional to the friction velocity squared. Hence:

$$u^* = \frac{u_z\ k}{\ln\left[\dfrac{z}{\dfrac{c\ u^{*2}}{2\ g}}\right]} \tag{3}$$

k is von Kármán's constant (0.4), u_z is the wind speed at height z and c varies from .026 over 70% snow covered lake-ice (Tabler, 1980) to 0.12 for snow covered grain fields (Pomeroy and Gray, 1990).

Suspended blowing snow derives from saltating snow and extends upwards from the top of the saltation layer via turbulent diffusion. One dimensional diffusion of sublimating particles undergoing gravitational settling in a turbulent atmosphere may be described as:

$$\frac{\partial\eta(z)}{\partial t} = \frac{-\partial\left(w(z)\,\eta(z) - K_s(z)\ \dfrac{\partial\eta}{\partial z}\right)}{\partial z} - c_{sub}(z)\,\eta(z) \tag{4}$$

where η is the mass concentration of snow in the atmosphere (kg m^{-3}), w is the mean vertical snow-particle velocity (m s^{-1}), κ_s is the turbulent diffusivity of blowing snow (m^2 s^{-1}) and c_{sub} is a sublimation rate coefficient (Pomeroy, 1988; 1989). For steady-state diffusion ($\partial\eta/\partial t = 0$), equation (4) may be solved as

$$d(\ln[\eta(z)]) = w^*(z)\, d\ln(z) \tag{5}$$

where w^* is a dimensionless diffusion parameter, lumping the effects of surface snow erosion, snowfall, sublimation, particle drag and atmospheric turbulence. Pomeroy (1989) suggests that snow diffusion proceeds from a reference mass concentration, $\eta(z_r)$, of 0.8 kg m^{-3}, whose height z_r is:

$$z_r = (0.05628\ s)\ u^* \tag{6}$$

For fully-developed blowing snow, he found the term w* varies with height as:

$$w^*(z) = (-0.8412\ m^{0.544})\ z^{-0.544} \tag{7}$$

Hence, one may find the drift density of blowing snow as a function of height and reference concentration height:

$$\eta(z) = \eta(z_r)\ e^{-1.55(z_r^{-0.544} - z^{-0.544})} \tag{8}$$

The transport rate of suspended blowing snow, Q_{susp}, is found by assuming the average horizontal particle speed approximates the average wind speed at that height and integrating from the lower reference, z_r, to the top of the blowing snow boundary layer, z_b:

$$Q_{susp} = \frac{u^*}{k} \int_{z_r}^{z_b} \eta(z)\ \ln\left(\frac{z}{z_0}\right) dz \tag{9}$$

where z_0 is the aerodynamic roughness height (m).

The vertical flux of snow at the top of the blowing snow boundary layer, q_v (kg m^{-2} s^{-1}) is solved from a steady-state mass balance of snow in a differential volume. Defining horizontal inflows equal to outflows, the upward vertical turbulent flux should be equal in magnitude to the mean downward settling flux of the snow particles. In mountain environments, particles that rise out the top of the boundary-layer are unlikely to return to it before sublimation or severe flow separation, hence:

$$q_v(z) = -w(z)\ \eta(z) \tag{10}$$

The mean vertical velocity, w, may be estimated from the dimensionless diffusion parameter (Shiotani and Arai, 1953) as $w \equiv w^*(u^*\ k)$ providing that the net effects from snow erosion and sublimation and dissimilarity between diffusion of momentum and snow particles are small:

$$w(z) = 0.3365\ u^*\ z^{-0.544} \tag{11}$$

This expression approximately agrees with profiles of terminal fall velocity calculated using Carrier's drag equation (Pomeroy and Male, 1986) and Schmidt's (1982) measured friction velocity and profiles of mean blowing snow particle size.

Sublimation of blowing snow can transfer significant quantities of snow to water vapour in relatively "mild" winter conditions. Sublimation is calculated using an energy-balance on individual particles (Schmidt, 1972), then integrating to ensembles of particles in realistic atmospheric boundary-layer conditions (Pomeroy and Male, 1987; Pomeroy, 1988). The important controlling factors are particle size, wind speed, air temperature and relative humidity. For instance the sublimation rate increases by one order of magnitude as the air temperature increases from -25° to -1°C or as the relative humidity decreases from 95% to 40%. Recent listings of the blowing snow sublimation equation and its application are given by Schmidt (1982), Pomeroy and Male (1987) and Pomeroy et al. (1991).

A simple model of blowing snow flow over wind-scoured terrain may be constructed by calculating the loss of snow due to sublimation within the boundary-layer and vertical transport from the boundary layer as a column of blowing snow traverses the wind-scoured zone. Noting that stream-wise velocity changes are assumed to be negligible and by definition of wind-scoured terrain, accumulation at the surface is nil, for a column of blowing snow extending from the surface to height z_b the change in transport rate with downwind fetch distance x is:

$$\frac{dQ_{0-z_b}}{dx}(x) = -q_v(x, z_b) - \int_0^{z_b} c_{sub}(x, z)\ \eta(x, z)\ dz \tag{12}$$

In considering changes in flow over wind-scoured terrain, $Q_{0-zb}(x=0)$ is the transport rate for fully-developed blowing snow.

THE CAIRNGORM EXPERIMENT

Because of persistently high wind speeds, heavy precipitation and a broad, yet irregular alpine/arctic setting the high plateaux of the Cairngorm Mountains, Scotland, are an ideal location to measure blowing snow. The profound redistribution of snow in these highlands is also extremely important to vegetation micro-climate, animal habitat, snow-melt runoff delivery, washout of snowpack chemicals, avalanche generation, ski-ing and other outdoor pursuits.

During March, 1989, vertical profiles of wind speed, snow particle flux, temperature and humidity were measured from a mast, located on a broad ridge connecting the peaks of Cairn Gorm and Cnap Coire na Spreidhe, above the corrie of Ciste Mhearad (see Fig. 1 for location). Field personnel remained on site each day, keeping instrumentation ice-free, oriented properly into the wind and elevated appropriately above the snow surface. Four "Vector" three-cup anemometers measured wind speed and four opto-electronic snow particle detectors measured particle flux within the lowest two metres of the atmosphere (Brown and Pomeroy, 1989). A datalogger recorded average values every five minutes.

THE BLOWING SNOW BOUNDARY-LAYER ON CAIRNGORM

Two days of contrasting measurements provide useful examples of conditions in the Cairngorms. The 22nd of March (snowfall) sustained winds over 70 km h^{-1} (at 10 m height), flowing along the ridge from Cairn Gorm to Cnap Coire na Spreidhe, and an approximate temperature of -6 °C and relative humidity of 85% during measurements. A continuous snow cover several tens of cm deep had formed on the ridge, supplemented by occasionally heavy snowfall during the day. The combination of strong winds and plentiful snow resulted in

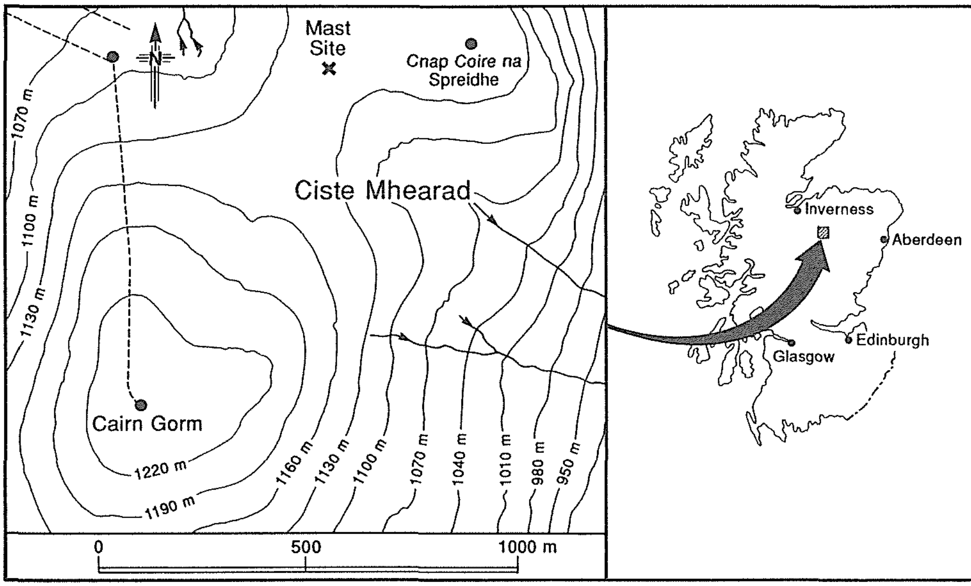

FIG. 1 Experimental site with meteorological mast on ridge above Ciste Mhearad in the Cairngorm Mountains, Scotland.

frequent white-outs. The 29th of March (wind-scour) sustained winds over 125 km h^{-1}, flowing up-slope from Ciste Mhearad and perpendicularly across the ridge, and an approximate temperature of -3°C and relative humidity of 90%. The ridge-top snow cover had melted into a sheet of boulder/gravel-strewn ice, whilst Ciste Mhearad maintained a substantial snow-cover, over three metres deep in places. The flow of wind-blown snow was funnelled on this day up out of Ciste Mhearad and onto the ridge between Cairn Gorm and Cnap Coire na Spreidhe; the surface as far as 700-1000 metres upwind of the mast remained essentially snow-free.

Representative five-minute profiles of wind speed are plotted in Fig. 2. The wind speed profiles are generally log-linear, indicating a fully-developed boundary-layer. Deviations from the log-linear case occur for strong winds above the one-metre height on 29 March,

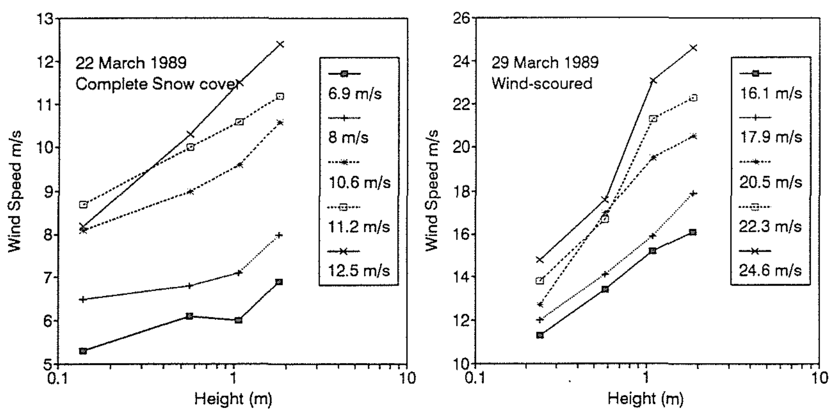

FIG. 2 Wind speed profiles on 22 & 29 March over complete snow cover and over wind-scoured terrain respectively. $U_{1.8\text{-m}}$ for the five-min averaged measurements is shown.

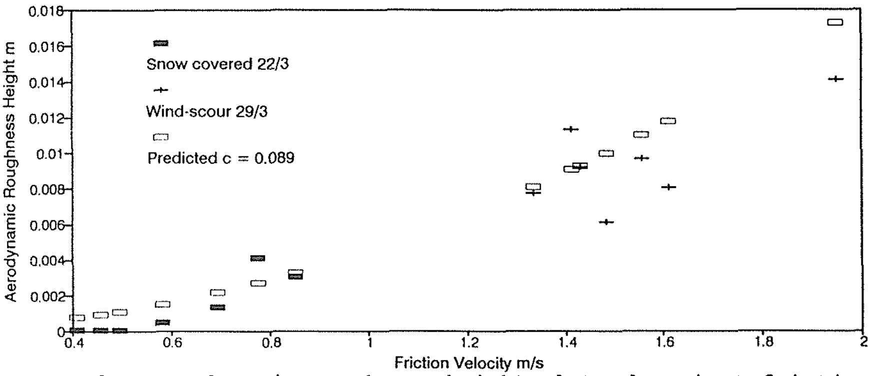

FIG. 3 Aerodynamic roughness height plotted against friction velocity, measurements and modelled. Values are 30-min or longer averages of uniform weather conditions.

indicating accelerations in flow of about 15%. These accelerations are of the order predicted using the irregular terrain flow model of Walmsley et al. (1989) set for flow over an appropriately scaled 2-dimensional ridge in rolling terrain. The relationship between aerodynamic roughness height and friction velocity was examined for time-averaged measurements (averages of periods with similar weather-type, typically 30 min.) and is plotted in Fig 3. Measurements on the 29th occurred during much higher wind speeds than on the 22nd making direct comparison between snowfall and wind-scour conditions difficult. However the coefficient, c, in Owen's (1964) saltation roughness relationship, $z_0 = c\, u_*^2/(2g)$, when set to c = 0.089 (fitted to a larger set of Cairngorm measurements) provides adequate simulation of the relationship for both snowfall and wind-scour.

In Fig. 4, profiles of measured blowing snow mass flux are plotted using the same five minute intervals shown in Fig. 2. The mass flux measurement heights are above the saltating layer, except possibly the lowest height on 29 March. The two conditions of measurement provide an interesting contrast; though wind speeds are twice as high on 29 March (wind-scour) the mass flux is up to an order of magnitude higher on 22 March (snowfall), an observation in conflict with equations (1) through (7). The measurements are compared to fully-developed mass fluxes, estimated from measured wind speeds, in Table 1. For the 6.9 m s^{-1} wind on 22 March measurements

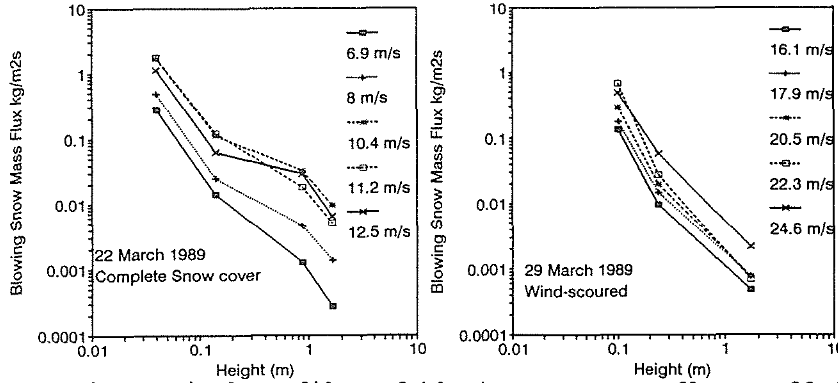

FIG. 4 Vertical profiles of blowing snow mass flux on 22 & 29 March over continuous snow cover with snowfall and over wind-scoured terrain respectively. $U_{1.8\text{-}m}$ for the five-min. averaged measurements is shown.

are well over fully-developed estimates due to an outburst of snow, falling and blowing from Cairn Gorm. Mass fluxes for the 12.5 m s^{-1} wind match the fully-developed case rather well when the vertical profile is considered, during this measurement a more moderate snowfall accompanied the blowing snow. On 29 March, mass fluxes are well below fully-developed values, ranging between 5 and 0.4% of full-development at heights above the saltation layer. The dramatic and consistent difference between these values and the fully-developed case indicates special consideration of flow over wind-scoured terrain is necessary to estimate blowing snow on 29 March.

TABLE 1 Examples of ratios of measured blowing snow mass flux to expected fully-developed mass flux.

	22 March			29 March	
Wind Speed at 1.8-m (m/s)	6.9	12.5		16.1	24.6
Height (m)	RATIOS		Height (m)	RATIOS	
.04	8.10	0.778	0.1	0.105	0.24
.14	26.2	0.588	0.24	0.0442	0.0285
.87	30.4	3.24			
1.67	12.1	1.04	1.72	0.0133	0.00417

MODELLING BLOWING SNOW OVER WIND-SCOURED TERRAIN

The wind-scoured condition in the Cairngorms on 29 March provides an opportunity to model the depletion of blowing snow transport and compare with measurements. Flow out of Ciste Mhearad provided an upwind wind-scoured fetch, with only slight distortion due to slope. The snow-covered fetch at the base of Ciste Mhearad was sufficiently long to permit fully-developed flow upwind of the wind-scoured fetch. Time-averaged (typically 30 min. of relatively uniform weather) wind speed, temperature and humidity measurements permit comparison of the measurements of mass flux at 1.72 m height with predictions; this height is used because lower measurement heights were near the height-variable saltation/suspension interface and provide difficult modelling for time-averaged measurements. Predictions were made using the fully-developed flow model, equations (1)-(7), and the wind-scoured flow model, equation (10). For the wind-scoured flow model, sublimation loss is calculated for z = 1.72 m. Vertical transport loss is calculated given the height z_b set to 2 m, and a proportional mass to that removed from the blowing snow column to satisfy vertical transport is removed from the horizontal mass flux at 1.72 m height. The time increment for flow at 0.15 m height is used, as mass flux profiles indicate this is a suitable "mass-weighted" average for the calculation. Two possible wind-scoured fetches are used, as shown in Fig. 5, the 1000 m fetch provides a better simulation than the 700 m fetch and a vastly better simulation than the fully-developed flow model. The accuracy of the "best" simulation suggests that the major processes contributing to snow transport rate decay over wind-scoured fetches have been accounted for. However higher-quality upwind meteorological data such as wind speed, temperature, humidity and mass flux at the beginning of the wind-scoured fetch or a less-simplistic simulation routine, such as an irregular topography plume-dispersion model may be required for improved accuracy of simulation. Note that whilst components of the wind-scour model could be "optimized" to provide a better fit than shown in Fig. 5, the limited array of field measurements does not

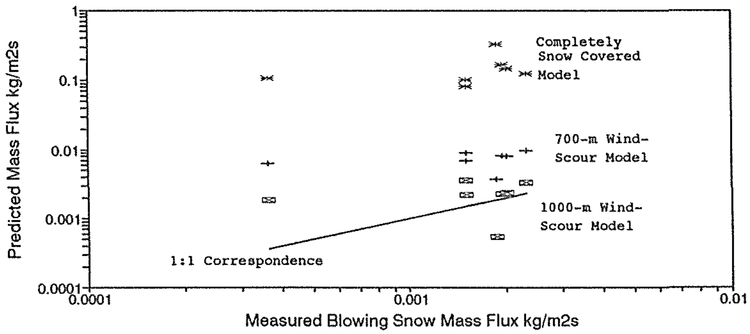

FIG. 5 Measured against predicted blowing snow mass flux for wind-scoured conditions. Models plotted are for fully-developed flow and for wind-scoured flow with 700-m & 1000-m fetch respectively.

justify changing model components until the phenomenon is more precisely understood.

Accepting the crude approximation of the wind-scour model, the downwind effect of sublimation and vertical transport losses on the transport rate may be indicated for one of the averaged conditions in Fig. 5. Using $u^* = 1.48$ m s^{-1}, $z_0 = 0.00617$ m, mean air temperature = -3.0°C and mean relative humidity = 88%, the downwind contribution of sublimation and vertical transport losses to the transport rate decay is plotted against fetch distance in Fig. 6. It is evident that for this relatively high humidity and high wind speed, vertical transport losses dominate over sublimation loss and that blowing snow transport will persist over a long wind-scoured fetch, though greatly reduced from its upwind value.

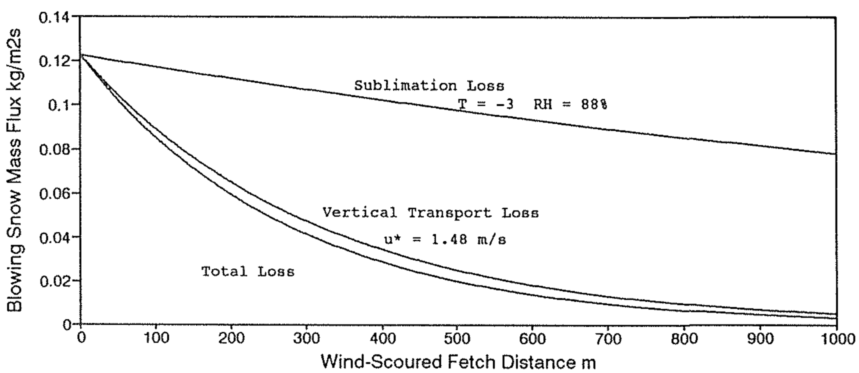

FIG. 6 Blowing snow transport rate decay along a wind-scoured fetch for u* = 1.48 m/s, temperature = -3°C, relative humidity = 88%, and decay due solely to sublimation and vertical loss.

CONCLUSIONS

a) The mass flux of blowing snow over wind-scoured terrain can be two orders of magnitude less than that over complete snow covers. Hence when an upwind supply of snow is transported over a wind-scoured fetch the supply arriving at the downwind edge is dramatically less than for a comparable snow covered fetch.
b) In high-wind-speed environments, the decay of blowing snow mass flux over wind-scoured fetches is largely due to vertical

transport of snow out of the surface boundary-layer. In mountainous terrain, short fetches make it unlikely that snow vertically removed to great heights can settle to the surface layer again before sublimating.

c) Sublimation of blowing snow particles that remain within the surface boundary-layer also contributes to the decay of blowing snow mass flux over wind-scoured terrain. Sublimation rates are extremely sensitive to humidity and temperature and hence vary widely with weather systems and local climate. Under relatively "arid" and "warm" winter conditions sublimation can contribute significantly to the decay of mass flux.

d) The dramatic losses of snow to vertical transport and sublimation indicated by the decrease in blowing snow over wind-scoured fetches suggest that a notable component of the winter's precipitation may be lost in certain alpine environments. The implications of these "open environment" snow processes should in particular be understood in regions where the extent of open areas is changing.

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