NUMERICAL SNOW DRIFT MODELING IN COMPLEX ALPINE TERRAIN AND COMPARISON WITH FIELD MEASUREMENTS

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ABSTRACT: Blowing and drifting snow influences human activities in various ways. For example, in mountainous regions, snow drift is a key factor in the formation of slab avalanches; the quality of avalanche danger mapping and land-use planning depends significantly on the correct assessment of snow redistribution by drift in avalanche release zones.

A physically-based numerical model was developed to simulate snow drift in alpine terrain. The model is based on the separation of the involved processes into two layers. The upper layer describes the transport in suspension and the wind field. The wind field is modeled by the Reynolds averaged Navier-Stokes equations, using the e–ϵ (k–ϵ) model for turbulent closure. Suspended snow is modeled by an additional scalar equation. The bottom layer describes the transport due to saltation, including erosion and deposition of snow. Both layers are mutually coupled by boundary conditions.

As an example, a numerical simulation for an Alpine ridge is presented and compared with field measurements. To this end, an experimental site 2 km north from the SFISAR Institute building at Weissfluhjoch, Gaudergrat ridge (2300 m a.s.l.), was equipped with instruments to measure meteorological and snow parameters.

The comparison between field measurements in a complex alpine terrain and numerical simulations shows that the snow drift model is suitable to predict the new-snow distribution in extended areas.

Keywords: snow drifting, numerical modeling, field measurements

1 INTRODUCTION

Snow drift is a key factor in the formation of avalanches as well as in the effectiveness of avalanches. Hence, great interest exists in the prediction of the influence of snow drift on avalanche danger. One can distinguish between the relevance for avalanche forecasting and land-use planning. In the former case, the primary interest focuses on the occurrence of snow drift and on the estimated enhancement of (new-)snow depths on a meso-scale (several 10 km^2) to improve avalanche warning on a regional scale. Usually, the periods of interest last for a few hours, but can also go on for days. In the second case, a consequence of the intensification in land-use is the increased importance of land-use planning. In alpine regions, avalanche danger is of major significance for the restriction of land-use, and great efforts are made to develop numerical avalanche models to determine endangered zones. These models require input data on the snow mass distribution in avalanche release zones (micro-scale; several 1000 m^2). Here, the interested depth of fracture is usually taken as the possible maximum increase of the snow depth during three days. Thus, there is a great interest for suitable tool to assess, analyze or forecast the effect of blowing and drifting snow in alpine topography.

Previous investigations of snow drift (and of the related sand drift) can be grouped into three categories: field measurements, physical simulations and numerical simulations. Most field work has been carried out on planes, e.g., in Antarctica (see Mellor and Fellers 1986), and concern mostly the determination of the transport rate for steady-state conditions, e.g., R.A. Schmidt (1986), or Kobayashi (1972). Only few attempts were made to measure snow drift in mountain areas, e.g., Schmidt et al. (1984), making drift flux measurements on a crest. Föhn and Meister (1983) looked into the snow distribution across a mountain crest. Physical simulations concern the deposition pattern around obstacles, e.g., Iversen (1980), or they deal with particular features of the drift process, e.g., the particle–bed impact as studied by a group around Maeno (1985) or by Rice et al. (1995). The existing numerical models of snow and sand drift differ widely in scope and focus. On one side, specific processes were studied, e.g., by Werner and Haff (1988) simulating the grain impact, but also models based on heuristic rules for avalanche forecasting, like ELSA (Mases et al., 1995) have been developed. At present, almost all models are

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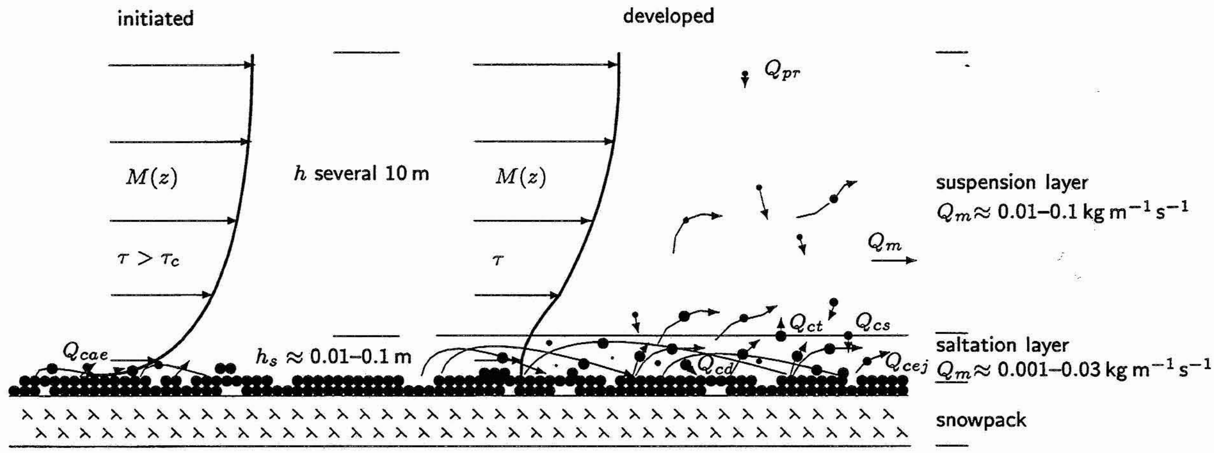


Figure 1: Schematic diagram of the processes involved in the drift system. The symbols denote: wind speed at a given height z, $M(z)$; shear/Reynolds stress, τ; critical shear stress, τ_c; aerodynamic entrainment, Q_{cae}; ejection rate, Q_{cej}; deposition rate, Q_{cd}; turbulent entrainment, Q_{ct}; settling rate due to gravity, Q_{cs}; horizontal mass transport rate, Q_m; precipitation rate, Q_{pr}. The acceleration of grains causes a modification of the wind profile.

one or two dimensional. Models based on empirical expressions for the saltation layer (Clappier and Castelle, 1991; Liston et al., 1993; Pomeroy, 1989; Uematsu, 1993), which describe the steady-state transport, are strictly speaking only valid for fetches long enough to achieve steady-state conditions. Treating erosion, most models do not distinguish between aerodynamic entrainment and ejecta due to impacts.

2 NUMERICAL MODEL

The transport due to blowing and drifting snow can be described as follows (see also Fig. 1). If the wind blowing over the snow surface becomes sufficiently strong, and wind shear exceeds a certain critical value, the so-called *threshold*, some grains are set in motion by the wind. Some of these grains will be lifted off the surface, and can be easily accelerated by the wind. Some of them will gain enough energy to rebound and/or eject other grains on impact. At the onset, the number of grains resulting from an impact, the so called *mean replacement capacity*, is, on average, larger than one. This results in an exponential increase of grains in the so-called saltation layer. Grains within the saltation layer follow more or less ballistic trajectories determined by the time-averaged wind profile and are not or only weakly influenced by turbulence of the wind. As more and more grains are in saltation, the vertical wind profile is modified due to the considerable extraction of momentum from the wind by the grains in motion. Now the grains gain less energy, and less grains will rebound or be ejected on impact. The mean replacement capacity decreases until the equilibrium value of one is reached. The number of saltating grains fluctuates around a certain value, sometimes called the *saturation value*. The surface properties, determining the threshold, as well as dislodgements effected by collisions of grains with the surface play an important role at the initiation of saltation. Thus, if there are already particles in the air, as is the case during snowfall, the threshold wind speed for drift is observed to be significantly reduced. Due to turbulent and gusty wind a certain number of particles go into suspension. In all cases, gravity acts as a back-driving force on the grains in saltation and suspension. Hence, snow drift is the result of five closely/mutually linked processes:

- aerodynamic entrainment,
- grain trajectories,
- grain–snowpack impacts,
- modification of the wind field,
- and the transport due to turbulent suspension.

When designing a numerical simulation scheme for blowing and drifting snow which can be solved with a reasonable computational effort, several constraints have to be considered. The major constraints will be the amount of involved snow grains and the spatial resolution of the two transport modes. For example, if we consider the saltation layer with a mass density of approximately $1\,\mathrm{kg\,m^{-3}}$

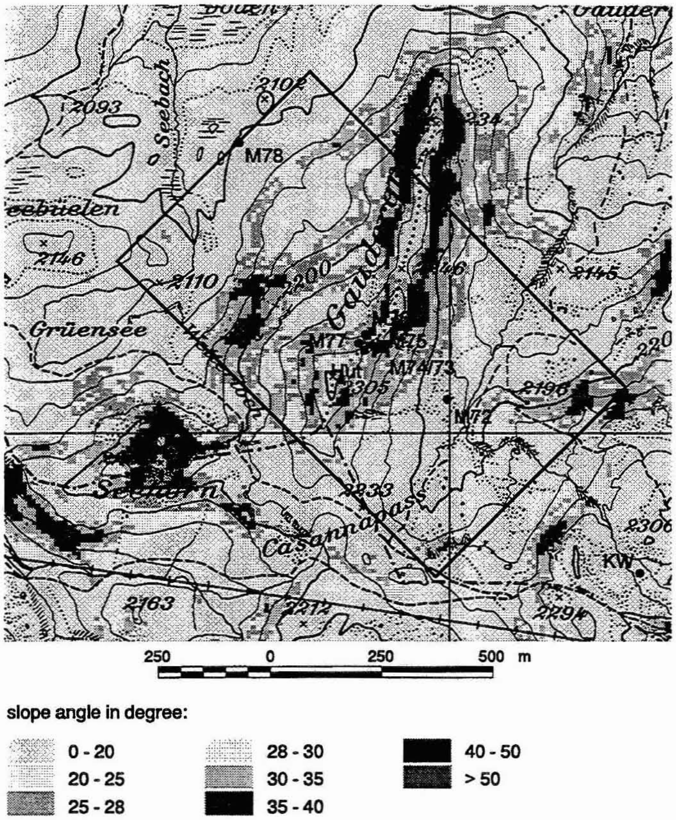


Figure 2: Slope angle map of the Gaudergrat experimental site showing locations of the masts Mxx and the hut that housed the transmitting equipment. The box marks the area of the numerical simulations. (Topographic data DHM25: ©Swiss Federal Office of Topography)

and a typical height of about 0.02 m, and if we assume a typical grain diameter of 200 μm, we get about $5 \cdot 10^6$ grains per square meter. Thus, particle dynamic methods, where the motion of single grains are calculated, lead for extended areas to computational demands which are beyond the computing power available today and the near future. Saltation with a vertical extension of about 0.01–0.1 m contributes significantly to the total transport. The transport due to saltation is in the range of 0.001–0.03 kg per meter width of the lane, per second, and starts at light winds. In comparison, suspension can extend to several 10 m in height, and the transport rates for the suspension mode range between 0.01–0.1 kg m^{-1} s^{-1} for moderate wind speeds. Resolving the saltation layer in a suitable way causes a drastical increase in the number of grid cells and a decrease of the allowable length of the time step. This leads to unacceptably long computing times.

To avoid these problems, a two-layer continuum model is proposed (Gauer, 1998). One layer describes the air flow and the transport due to suspension. The core of the suspension-layer modeling constitutes a simple mixture formulation, which is based on the equation of state, the conservation equations for mass and momentum, and an additional conservation equation for moisture. For the turbulent closure, the e–ϵ model is used. Within the second layer, the height averaged transport due to saltating grains is solved. Here, the starting point is formed by the mass conservation for the mixture of air and snow, and the mixture momentum equation which describes the balance of the force necessary to accelerate the saltating particles and the driving forces, represented by the turbulent shear stress and gravity. Both layers are mutually coupled by boundary conditions describing the snow mass exchange, the wind speed as well as the turbulent shear stress at the top of the saltation layer. The boundary conditions between the saltation layer and the snowpack are determined by the aeordynamical entrainment of snow grains, grain ejecta due to impacts and grain deposition.

3 FIELD MEASUREMENTS AND COMPARISON WITH SIMULATIONS

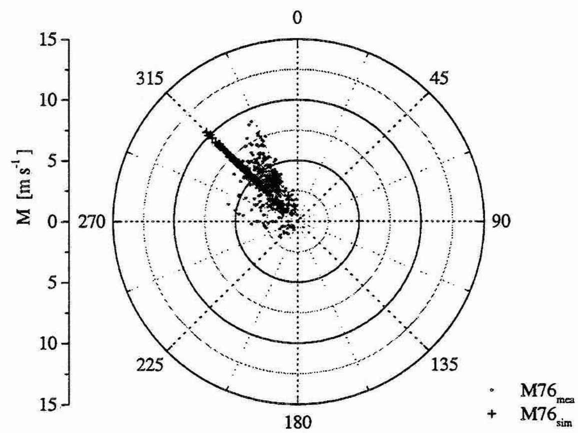
For a validation of numerical snow drift models, two points are of significance: 1) the meteorological input, particularly the wind field as the driving force for blowing and drifting snow, and 2) quantities describing the drift, e.g., flux measurements during drift or measurements of the areal snow redistribution pattern for a drift episode.

Field measurements of drifting snow in complex terrain are rather scarce, and no example is known where simultaneous measurements of both, the wind field and the snow distribution, had been carried out.

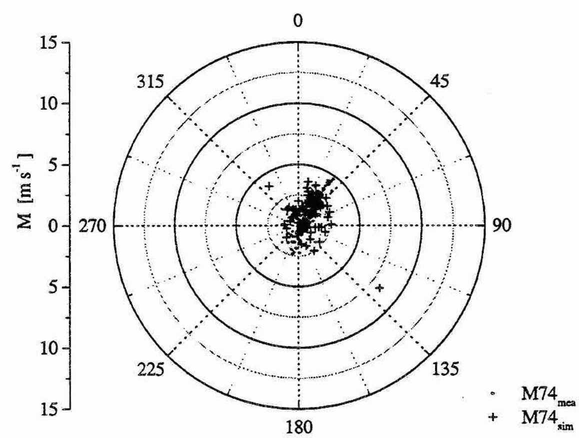
With this in view, measurements were carried out during the winter 1996/97 in complex alpine terrain. Gaudergrat ridge, 2 km north of Weissfluhjoch/Davos was chosen as field site and equipped with instruments to measure meteorological and snow parameters. The ridge has a rather sharp crest—the slope angles range from 28 to 38°—and might be regarded as prototypical of Alpine topography. The prevailing wind direction during strong precipitation periods is more or less north-west and thus perpendicular to the crest-line. In order to determine the wind field around the crest, the site was instrumented with five masts in the surrounding area for wind profile measurements. A sixth mast was equipped with additional sensors for measurements of meteorological and snowpack parameters. Figure 2 shows the locations of the installed masts and the slope angles at the ridge.

To determine the areal snow redistribution for a drift episode, soundings of the snow depth were made before and after an episode, that is, the snow depth was measured along equidistant lines across the crest, roughly 8.5 m apart and over 200 m long. The soundings were taken at 4 m intervals along the lines; in the neighborhood of the crest this distance was reduced. Thus, every field campaign resulted in around 350 data points. All measurement points were marked by thin bamboo stakes during the first campaign. Thus, later measurements could be taken at the same points (with an estimated horizontal deviation of less than 0.1 m). In this manner uncertainties in the depth measurements due to the small-scale topography could be minimized.

During winter 1996/97, six snow drift episodes could be investigated. An example of wind field measurements and the corresponding numerical simulations is given in Figure 3. The figure shows two polar plots combining the measured and simulated wind speeds for drift period of 26–27 February 1997. Generally, the simulations agree quite well with the measurements. While the simulated wind speeds are too high at the crest (M75) they tend to be too low at the outflow of the domain (M72) (both are not shown here). Nevertheless, the simulated wind field gives reasonable results, particularly if we keep in mind that boundary conditions for this complex terrain are poorly known.



3.a: windward side of the crest



3.b: leeward side of the crest

Figure 3: Comparison of simulated (+ ; 25 min averages) and measured (· ; 5 min averages) wind speeds and directions. Polar plots corresponding to two sensor masts.

Figure 4 depicts an example of the simulated redistribution pattern for the new-snow layer. The new-snow depth is normalized against the reference depth, HN_{ref}, which corresponds to the new-snow depth in an undisturbed area. The new snow is picked up in acceleration regions such as the area close to the crest line, at small humps and brows. At the crest line, erosion of the old snow pack is observed as well. Deposition occurs in the

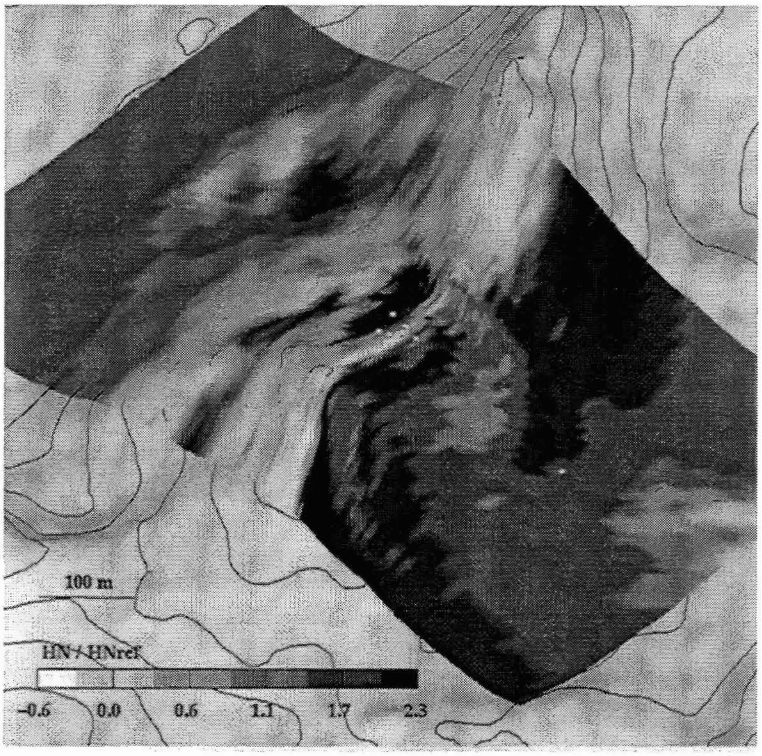


Figure 4: Simulated snow distribution. The snow depth is normalized by the reference new-snow depth; negative values mean erosion of the old snow pack. (HN_{ref} = 0.2 m; after t = 35 h)

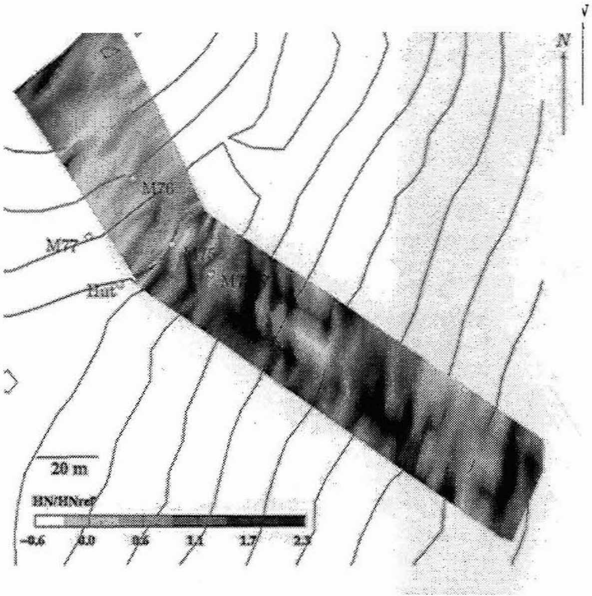
deceleration regions leeward of the saddle, in small gullies and in hollows. In this simulation deposits of twice the reference height and erosion of about 0.1 m of the old snowpack at the crest line are found.

A comparison between soundings taken at the experimental site and the simulation is shown in Figure 5. The measurements show great variability, but also some trends can be discerned: the erosion area in the windward slope near the crest line, the formation of a cornice at the ridge, and an area just leeward of the crest where the snow depth is less than the reference depth. Down-slope areas where HN is twice HN_{ref} alternate with depletion areas such as the small terrace on the right. It should be pointed out that the small-scale terrain features in combination with the high turbulent wind cause large differences in the erosion and deposition patterns. The simulation does not show such great variability, but nevertheless reproduces some characteristic features like the windward erosion area close to crest, and alternation between depleted areas and areas of enhanced deposition leewards. One reason for the differences between the measurements and the simulations is surely the restricted resolution of the local topography used for the simulations.

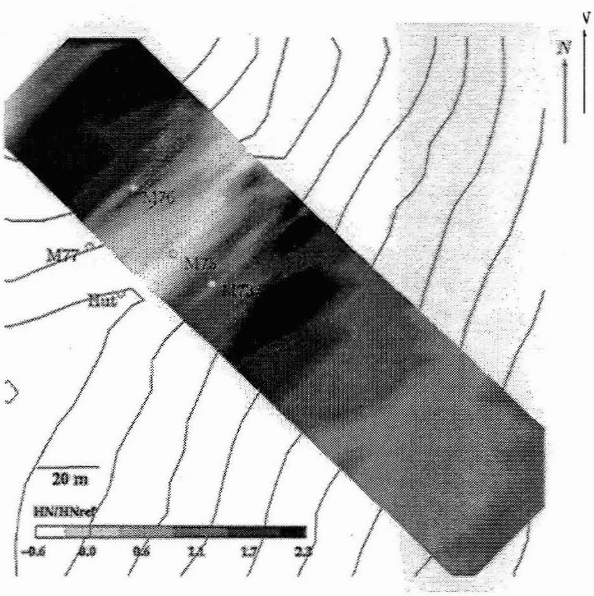
Figure 6.b shows a comparison between the measured and the simulated temporal evolution of the new-snow depth at points M77 and M74 (windward and leeward of the crest; both mast were equipped with ultra sonic snow depth sensors). The comparison shows that the simulation reproduces the temporal evolution of the snow depth quite well. In addition, Figure 6.a shows the corresponding wind speed measurements at the top of the crest and the measured precipitation rate. At the top of the crest, the wind direction was more less north-west during the whole time. Out of Fig. 6.b and Fig. 6.a, it is obvious that the onset of a strong windward erosion starts as the wind speed exceeds a certain wind speed. For this episode, the estimated critical shear stress, τ_c, was in the range 0.2–0.4 Pa.

4 CONCLUDING REMARKS

The proposed model is a step towards numerical simulation of blowing and drifting snow in alpine terrain. The primary emphasis is on a physically-based description of snow drift suitable for complex alpine terrain. The model includes: a fully 3-dimensional non-steady-state modeling; the modeling of the two main transport modes (saltation and suspension); the dynamical modeling of deposition and erosion, distinguishing between aerodynamic entrainment and ejecta due to particle

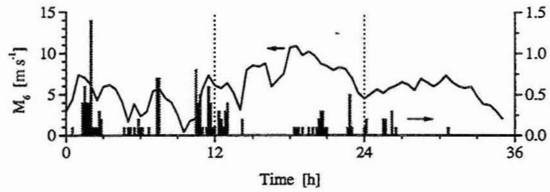


5.a: Measured new-snow depths

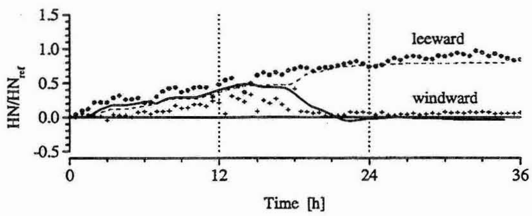


5.b: Simulation ($HN_{ref} \approx 0.2$ m; after t $\approx$ 35 h)

Figure 5: Comparison between soundings and simulated redistribution pattern for a small area around Gaudergrat ridge. Shown are the normalized new-snow depths; negative values mean erosion of the old snowpack. (Contour interval is 10 m)



6.a: Wind speed, $M(z = 6\,\mathrm{m})$ (30 min averages), at the top of the crest and the precipitation rate, RS (10 min sum); wind direction, dd $\approx$ 315° (north-west).



6.b: Temporal evolution of the normalized new-snow depth at the points M77 (windward) and M74 (leeward). (Lines denote simulation and dots are measurements)

impacts; the back-reaction of the particles in saltation on the flow (two-way coupling).
The comparison between simulations and field measurements gave encouraging results that numerical simulations of snow drift will develop into a powerful tool for land-use planning and avalanche forecasting.

On the other hand, the first numerical simulations also demonstrated the problems involved in the simulations of complex terrains. For example, the grid resolution must match the length scale of the terrain and the wind field at the boundary of the computational domain needs to be interpolated correctly. The former requires input data of the topography with a high resolution. To overcome the problem with the poorly known boundary conditions, one might consider the adaptation of the output of a meso-scale weather model. There are some of those models in developing, but not yet fully adapted to use in mountain areas.

A critical point in snow drift simulation is the nonlinear increase of the computational effort with increase of the area of interest and/or spatial resolution of the area. At that point, a compromise between a high spatial resolution and an acceptable computational effort has to bee found.

A great unknown is the influence of the snowpack on snow drift. The snowpack properties determine parameters like the erodibility, the rebound probability of particles etc. Little is known about the coupling between wind and natural snow. The parameterizing for erosion and deposition used in the presented model is mainly based on wind tunnel studies and numerical simulations for aeolian sand transport. Some research will be necessary to achieve better adaptation to the properties of snow. In a further step, one can aspire a coupling between snowcover models and snowdrift models.

ACKNOWLEDGMENTS

I like to thank D. Issler and P. Föhn for many enjoyable discussions, and B. Brown for his comments. I am also indebted to W. Ammann, C. Fierz, M. Hiller, G. Klausegger, G. Krüsi, R. Meister, and P. Weilenmann, as well as the staff of the mechanical work shop for the help in preparing the field measurements. Similarly, I am grateful to C. Camponovo, S. Harvey, C. Marty, R. Mullins, M. Phillips and K. Plattner for their assistance in the field, and A. Stoffel for his GIS support.

This work was supported by the Swiss Federal Office of Education and Science under the contract OFES 93.0082 as part of the European "Human Capital and Mobility" project "Contribution to avalanche dynamics with the aim of mapping hazardous areas", contract CE CHRX-CT 93-0307.

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