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Assessment of snow transport in avalanche terrain

Michael Lehning *, Charles Fierz

WSL, Swiss Federal Institute for Snow and Avalanche Research SLF, Davos, Switzerland

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Abstract

A local to regional assessment of transported snow during snow storms or subsequent periods of strong winds is a prerequisite to reliably estimate avalanche danger. Despite the fact that it has received continuing attention for decades, the problem of quantifying snow transport persists. Systems from point measurements to full three-dimensional simulations have been tested but all have their respective weaknesses. We present a new drift index, which has been tested and operated with some success in Switzerland. The index requires input from a wind-sheltered automatic weather station and a scaled wind speed from a wind-exposed site. Using the snow cover model SNOWPACK, the meteorological data is extrapolated to the four main aspects and snow cover development is calculated for these aspects. Depending on the measured wind direction and speed, a threshold condition for snow erosion at the upwind aspect is tested: if the wind is strong enough to erode the current snow at the surface of this aspect, the snow layer is eroded, transported and deposited onto the downwind aspect. With this scheme, the virtual, "representative" snow cover on the four main aspects in the vicinity of the meteorological stations are reconstructed for the course of the winter and the mass transport rate is converted to a lee-deposition drift index. A comparison with FlowCapt, an acoustic measurement device, which measures a local mass flux, shows that the measured mass flux correlates well with the amount of lee-slope deposition predicted by the drift index. Also, drifting snow periods are well detected by both the FlowCapt sensor and the SNOWPACK drift index and correspond to drifting snow periods reported by local observers. When comparing regional patterns of strong and weak snow transport as calculated from more than 110 automatic weather stations in the Swiss Alps with corresponding reports from local observers a good correlation is found, too. As opposed to earlier versions of the index, which had been based on flat field simulations of SNOWPACK alone, the new index no longer overestimates intensity and duration of blowing snow events. It is concluded that for the purpose of avalanche warning, the FlowCapt sensor and the SNOWPACK drift index are suitable means to quantify local to regional snow transport.

Keywords: Drift index; Lee slope loading; Avalanche warning; SNOWPACK; Snow cover simulation

1. Introduction

Snow transport is a major factor in determining the local avalanche danger. For the purpose of avalanche warning, snow transport is usually assessed by visual inspection of blowing snow clouds over mountain ridges or subjective observations of snow drifts. A few measurement devices have been designed (e.g. FlowCapt, Chritin et al., 1999), which measure the transport rate at a single point. They are not widely used because a point transport rate is difficult to interpret and the devices have had calibration problems (Lehning et al., 2002). Therefore, a different approach has been proposed which is to use model descriptions of drifting and blowing snow to

* Corresponding author. Snow and Avalanche Research SLF Flüelastr. 11, CH-7260 Davos Dorf, Switzerland.
E-mail address: lehning@slf.ch (M. Lehning).

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describe snow transport over mountain ridges and loading of lee slopes by wind transported snow. While first attempts have been made to model the full three-dimensional distribution of snow deposition and erosion on a scale relevant for avalanche warning (Gauer, 2001), for operational purposes, a simplified approach is necessary. To our knowledge, the only two operational model systems that can include snow transport assessment are the French Safran–Crocus–Mépra chain with the SYTRON blowing snow model (Durand et al., 2004) and the Swiss snow drift index (DI) based on SNOW PACK model simulations (Lehning et al., 2000). However, due to systematic weaknesses, the SNOWPACK drift index has had limited success and acceptance in the avalanche warning community. We therefore have continuously worked to improve the index and present now a new DI algorithm, which is still based on measured wind and simulated snow cover properties. In the following we discuss the index improvements, in particular the introduction of virtual slopes for four aspects and the scaling of the measured wind speed. We compare the new index against observations of snow drifts from local observers at individual locations. We also compare the SNOWPACK DI against the measured mass fluxes of the acoustic sensor FlowCapt. Finally, the distribution of drifts observed by local observers is compared against the snow transport pattern as predicted by the DI.

2. Drift index algorithm and set-up

The mass transport calculation follows the algorithm presented in Lehning et al. (2000). First, a local threshold friction velocity, u_{*th}, is determined based on the local snow characteristics as calculated by the snow cover model SNOWPACK (Lehning et al., 1999):

$$u_{*th} = \sqrt{\frac{A\rho_i g r_g (SP+1) + B\sigma N_3 \frac{r_b^2}{r_g^2}}{\rho_a}} \qquad (1)$$

where ρ_i and ρ_a are the ice and air density, respectively, SP is the sphericity of the model grains, r_g and r_b are the grain radius and bond radius, respectively, N_3 is the grain coordination number, g is the acceleration due to gravity and σ is a reference shear strength set to 300 Pa. Eq. (1) is theoretically derived from a force and overturning moment balance considering the weight of a particle to be lifted from the surface (term 1) and the force needed to break the bonding to neighboring grains (term 2) for a certain assumed geometry (Schmidt, 1980). Our wind tunnel verification studies (Clifton et al., 2006) and field investigations have shown that the two geometrical constants A and B need to be set to 0.02 and 0.0015 respectively. All the other parameters are snow microstructure parameters calculated by SNOWPACK for the individual snow layers. The resulting threshold friction velocity is a measure of the drag that the atmosphere needs to exert on the snow surface for drift to start. In earlier versions, the DI was generally overestimating real drift. One reason was that the original constants in Eq. (1) caused the resulting threshold friction velocities to be too low. Using the threshold friction velocity, a saltation mass transport rate, Q (kg m^{-1} s^{-1}), can be theoretically predicted from boundary layer theory (Schmidt,1986). This basic concept, which looks at the excess shear stress above the threshold has been altered to fit measurements (Sørenson, 1991):

$$Q = 0.0014\rho_a u_* (u_* - u_{*th})(u_* + 7.6u_{*th} + 205), \qquad (2)$$

where u_* is the friction velocity derived from the measured wind speed assuming a local roughness and a neutral logarithmic velocity profile. For a given roughness and therefore surface, the friction velocity is proportional to the wind speed. For typical surfaces in Alpine terrain, friction velocities translate to 2 m wind speeds with a factor between 10 and 20. Assuming a typical roughness, which can be described with a roughness length of 2 mm, a 2 m wind speed of 5.1 m s^{-1} has an associated friction velocity of 0.3 m s^{-1}, which could be a typical threshold for drift over a fresh snow surface.

This saltation model is used here as it has proven to deliver good results when compared to measurements or more physically based models (Doorschot and Lehning, 2002). It now replaces the earlier implementation following Pomeroy and Gray (1990), which gives a linear dependence on friction velocity, while it is known that saltation mass flux has a power law dependence on local wind speed with an exponent between three and four (e.g. Nemoto and Nishimura, 2001). The local snow mass transport rate is then translated into a lee-slope drift, by assuming that the extra mass is distributed over a slope length of 70 m. This arbitrary slope length is only used to translate the local transport rate to a hypothetical extra deposition in the lee slope due to drifting snow, which is easier to interpret for avalanche practitioners. It can be considered a scaling parameter without physical meaning. Note that prediction of transport in saltation is taken here as a proxy for total transport, since transport in suspension cannot be adequately represented by a one-dimensional model.

By now the operation of the SNOWPACK model in the Swiss Alps is based on more than 110 automatic weather stations (AWS), the altitude of which ranges from 1560 to 3345 m a.s.l. Each AWS consists of a wind and one to three snow stations. The wind station is at a ridge or summit and the snow station(s) is (are) at representative flat-field site(s) close to the wind station. The measured wind at the wind station is combined with the simulated snow at the snow station to calculate a virtual mass transport rate as described above and an associated lee slope loading. This combination has proven to be problematic because the snow station is chosen to be at a representative snow site, where snow deposition or erosion due to wind should not be dominant. Therefore, the snow at the snow station is in reality never exposed to the high wind speeds measured at the wind station. Accordingly, drifting and blowing snow does not influence the surface of the modeled snow and an undisturbed surface is exposed, which in turn is used to calculate the threshold wind speed as discussed above. This leads to an overestimation of the duration and strength of modeled blowing snow events while in reality all moveable snow may have already been eroded or compacted by the wind.

The new drift index algorithm solves this problem by introducing four virtual slope aspects with a slope angle of 38°. For the four aspects north, east, south and west, separate snow cover simulations are introduced. Per calculation time step the following algorithm is applied:

1) From the measured wind direction at the wind station, the windward and lee aspects are determined.
2) The threshold wind speed for the windward aspect is determined.
3) If the measured wind is higher than the threshold, the mass transport rate is calculated as explained above.
4) The mass transport rate then determines the amount of snow eroded at the windward aspect.
5) If a complete SNOWPACK snow layer is eroded, the threshold conditions for the newly exposed snow layer are re-evaluated not before the next time step.
6) The eroded mass and the density of the eroded layer are stored and then deposited at the lee aspect.

Note that snow from snowfall is equally deposited at all aspects and the flat field. Since freshly deposited snow is very easily eroded, this leads to the fact that during snow fall with moderate to high wind speeds snow transport rates are usually high. Finally, snow transport rates are averaged over 6 and 24 h irrespective of aspect to yield a representative snow drift index. In the analysis, we neglect sublimation of airborne snow, which is consistent with the aim of constructing a simple index, but may lead to an overestimation of drift snow in case of special conditions such as a Föhn storm, when high local transport rates only lead to minor snow loading in lee slopes.

3. Normalization of wind speed

A complication with the combination of wind and snow stations as described above is that the relationships between wind and snow stations are very site-specific. Since the snow transport rate in saltation is approximately a function of the wind speed to the third or fourth power (see Section 2), small changes in the wind speed will cause large changes in the snow mass transport rate. As a result, plotting the simulated DI on a map of the Swiss Alps, very inhomogeneous patterns will result, depending more on the wind statistics of the individual wind stations than on the true local to regional snow transport. Therefore, we need to normalize the wind speed measured at the wind station to allow individual stations to have comparable drift index values for comparable conditions. A normalization based on the mean and the standard deviation has proven to be effective: For every station, a correction factor, f_s, is calculated based on a time series of at least one complete winter season:

$$f_s = \frac{\frac{1}{n}\sum_{i=1}^{n} (\overline{v}_i + \sigma_i)}{(\overline{v}_s + \sigma_s)} \tag{3}$$

where n is the number of stations in a region, the subscript $_s$ stands for a particular station, $\overline{v}$ is the time-average of the wind speed measurements and σ the corresponding standard deviation. Values of f_s have been found to vary between 0.7 and 1.4 for ridge wind stations, with many values being close to unity. The simple sum of mean and standard deviation leads to an adequate scaling of stations with a high gustiness of the wind, which would result in very high transport rates because of the non-linear dependence on the wind speed as expressed in Eq. (2).

4. Validation and discussion

The validation of the DI as described above is not a trivial task. This is because there exist no reference measurement methods of neither local nor regional snow transport in alpine terrain and we are left with indirect validations. We suggest four independent indirect validations, which give in combination a sufficient picture of the performance of the DI. The four comparisons are:

1) Snow re-distribution and deposition over a single mountain ridge for individual storm events.

2) Comparison of predicted blowing and drifting snow events at single IMIS stations over a full winter season with reports from observers in the vicinity.
3) Validation of drift index prediction at single IMIS stations against FlowCapt measured snow transport rate at the same IMIS (snow) station.
4) Comparison of regional or mountain scale patterns as predicted by the SNOWPACK DI with patterns arising from our observer network.

In the following we will discuss the single validation attempts in more detail. While we present new material for the latter three points, we only review older information for the first point.

4.1. Comparison of single storm, single mountain transported mass

Our earlier investigations have shown that calculations with a physically based saltation model and a simplified suspension model, which takes the deviation of the mean wind profile over a mountain ridge into account, yield very good estimates of the total mass transported over a mountain ridge during a snow storm (Doorschot et al., 2001). This work compared transported mass as measured by manual snow deposition measurements in the lee slope with predictions from the blowing and drifting snow model for four storm events in the years 1997 and 1999. Without changing the model set-up or parameterizations, very good predictions of total snow transport have been achieved for all snow drift periods at the same ridge. This result is one important verification of the saltation model of Doorschot (Doorschot and Lehning, 2002), which has been shown to produce comparable results to the empirical saltation formula of Sørenson (Eq. (2)). Accordingly, the latter is used here for computational efficiency. Since the changing snow surface conditions have a very small influence on total snow transport for these short-term storm events, the results can only be regarded as a partial validation of the DI presented here. Furthermore, local measurements of wind speed at the ridge and at the inflow have been used to calculate the mass transport rates, taking into account the speed-up profile over the ridge and its influence on snow transport. This is not the same set-up as for the DI discussed here and therefore additional verification studies are necessary.

4.2. Comparison against local observational judgment of snow drift periods

In this section, we compare drift index calculations with observations of drifting snow made by local observers. Note that the observations are usually taken from an observer in the vicinity of the automatic weather station. The observers are based at valley stations and report the occurrence of drifts in four classes: class 0 — “no reported drift” or up to 5 cm, class 1 — “5 to 20 cm”, class 2 — “20 to 50 cm” and class 3 — “larger than 50 cm”. Below we make simple binary (drift/no drift) comparisons as well as a verification, for which the quantitative drift index calculations have also been reduced to the same classes.

Fig. 1 shows the comparison between the new DI described here and the original version (Lehning et al.,

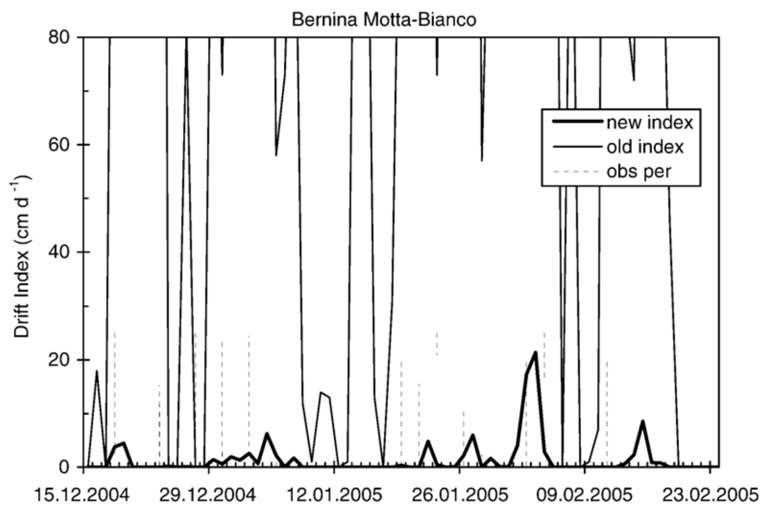

Fig. 1. Drift index from the AWS Bernina Motta-Bianco during the winter 2004/2005 for the old and the new SNOWPACK drift index, thick and thin solid line, respectively. The light dotted line shows times when drifting snow was reported by local observers.

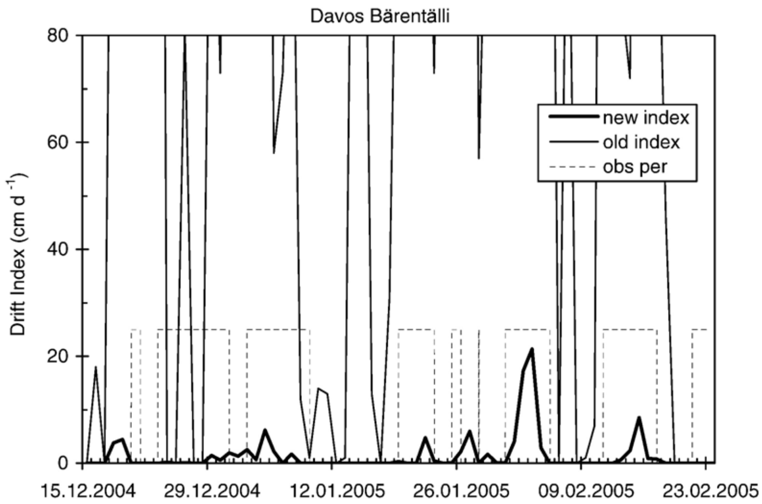

Fig. 2. Drift index from the AWS Davos Bärentälli during the winter 2004/2005 for the old and the new SNOWPACK drift index, thick and thin solid line, respectively. The light dotted line shows times when drifting snow was reported by local observers.

2000), exemplified at the automatic weather station Bernina Motta-Bianco in the south of the Alps in the canton Grisons. This is an extreme example in the sense that the wind station Bernina Lagalb at an altitude of 2959 m a.s.l. has very high average wind speeds. Also, the snow station, Bernina Motta-Bianco is at 2450 m a.s.l. more than 500 m lower and will therefore have a very different snow cover. Therefore, the old, un-scaled DI gave always unreasonably high values because of the un-scaled wind but also because of the conceptual problems and the threshold problems discussed above. The wind speed scaling factor (Eq. (3)) for the station is $f_s = 0.8$. The non-linear effect on the mass transport is clearly visible in that reducing the wind speed by 20%, using improved threshold friction velocities and introducing the slope simulations, much lower DI values result. The maximum 24 hour DI has been reduced from 400 cm to 30 cm. The new index gives reasonable values of estimated snow transport and the shorter periods of simulated snow drift correspond better to the corresponding estimates of drift periods from our local experts (light dotted line in Figs. 1–4). Note that observers tend to report drifts for even minor indications of potential drifting snow such as a small plume of blowing snow over a ridge.

Fig. 2 shows the same comparison as Fig. 1 but for the IMIS station Davos Bärentälli (2560 m a.s.l.) in the central eastern Swiss Alps, which uses the wind measured at Krachenhorn (2891 m a.s.l.). Again, the old index produced far too high drift index values up to

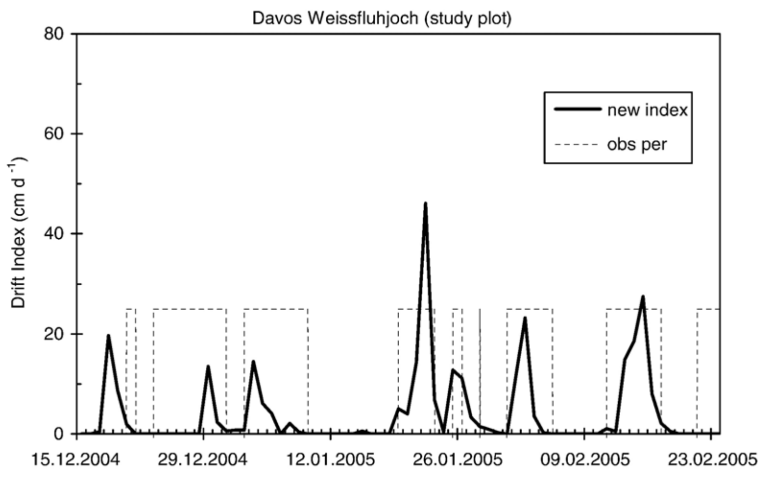

Fig. 3. New drift index from the AWS Davos Weissfluhjoch (study plot) during the winter 2004/2005. The light dotted line shows times when drifting snow was reported by local observers.

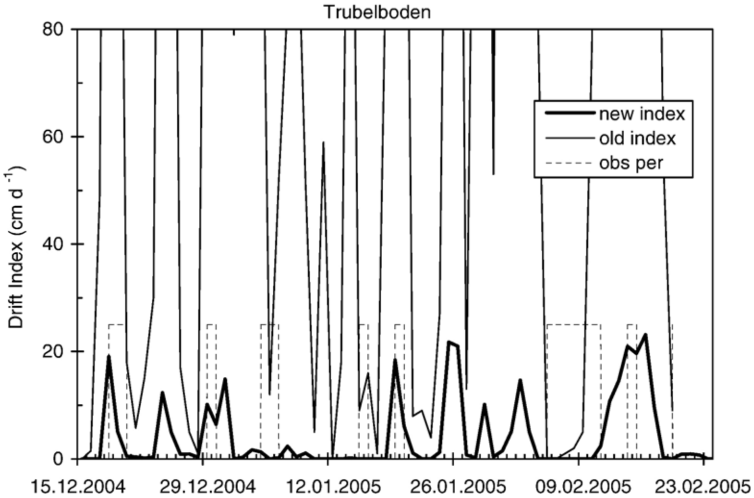

Fig. 4. Drift index from the AWS Trubelboden during the winter 2004/2005 for the old and the new SNOWPACK drift index, thick and thin solid line, respectively. The light dotted line shows times when drifting snow was reported by local observers.

120 cm. This is, however, purely caused by the threshold and conceptual problems of the old index, since the scaling factor (Eq. (3)), f_s, has been calculated to be 1.19 for this station, which means that at Krachenhorn below-average wind speeds are measured. For both stations, Bernina and Davos Bärentälli, the observed snow drift periods correspond well to the modeled ones.

Fig. 3 shows the comparison of observed and modeled (new DI only) snow drift periods for the study plot Davos Weissfluhjoch (2540 m a.s.l.) based on wind measurements at the Weissfluhjoch (2693 m a.s.l.). Here, larger DI values of up to 50 cm are calculated for approximately the same snow drift periods as for Krachenhorn. This can be regarded as a typical local to regional variation, as Krachenhorn is located approximately 17 km Southeast of Weissfluhjoch. Note that the observed snow drift periods are identical as they are from the same expert in the valley.

A further example of observed versus modeled snow drift is Fig. 4 for the station Trubelboden (2480 m a.s.l.) in the middle part of the canton Valais. Here the corresponding wind station is at 3096 m a.s.l. and maximum drift values observed with the old DI are as high as 337 cm. For this example, the wind speeds are now also reduced by the scaling factor ($f_s = 0.82$) and more reasonable drift values with a good correspondence to the observed snow drift periods result. An exception is the period at the end of January/beginning of February, when the correlation between model and observation is poor. A possible explanation could be the distance between the observer and the automatic weather station, which is approximately 20 km in this case.

If we classify DI values according to the observer classes, we can make a semi-quantitative verification. Fig. 5 shows this comparison for the AWS Davos Versuchsfeld Weissfluhjoch. In 46% of all cases, the drift class has the same value both in observation and simulation. In 11% of all cases drift was both observed and simulated but the classes differed. This gives a "total success rate" of 57%. Overall the DI appears to underestimate drift for this case. The same analysis was done for five automatic weather stations and total success rates ranged from 55% to 73%. Higher success rates cannot be expected since an observation in a valley and a drift index calculation for a high mountain AWS are not identical.

4.3. Verification against FlowCapt snow drift sensor measurements

FlowCapt (Fig. 6) is an acoustic sensor, which analyses the sound of the impact of snow particles on a tube with built-in microphones (Chritin et al., 1999). An analysis of early versions of the sensor (Lehning et al., 2002) has shown that the sensor could be useful for snow drift monitoring at IMIS stations, despite the fact that rather severe calibration problems were present. In the meantime, the calibration of the sensor has been improved and a few IMIS stations have been equipped with FlowCapt sensors for the assessment of local snow transport flux. The IMIS FlowCapt sensors consist of three segments with three separate microphones. Each segment is one meter high and nominally the lowest segment measures saltation and the two upper segments

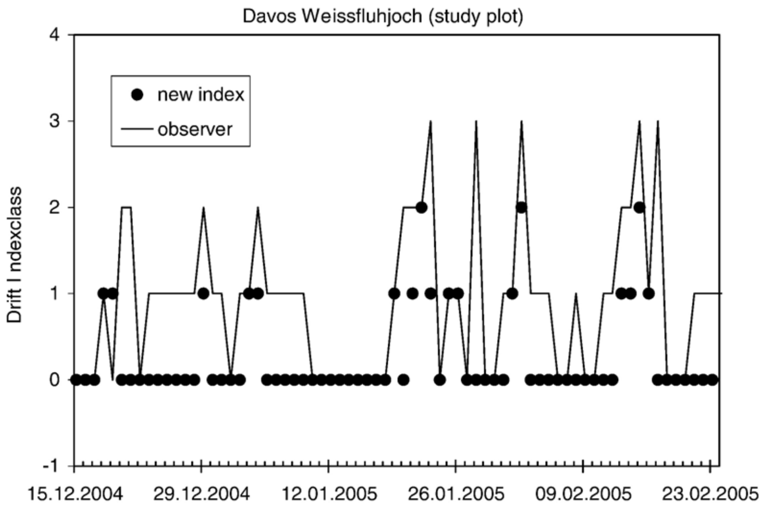

Fig. 5. New drift index and observed drift classified in four classes (0 to 3) for the AWS Davos Weissfluhjoch (study plot) during the winter 2004/2005.

measure suspension mass flux. For the comparison described here, the average flux measured over the three segments (g m^{-2} s^{-1}) is presented. Note that the SNOWPACK DI analyzed here gives lee slope loading for the past 24 h at a time resolution of 3 h. Therefore, the FlowCapt measurements have also been averaged (running average) over 24 h and sampled every 3 h to allow a direct comparison with the DI. Since the mass flux is usually decaying exponentially with height (Clifton et al., 2006), the FlowCapt measurements have also to be regarded as an index measurement rather than a precise flux measurement. Problems that have been observed with the FlowCapt sensor include riming, erroneous drift readings due to wind noise and problems due to local snow accumulation.

In the following, we present a direct comparison of measured local transport rate and the SNOWPACK DI prediction. Since for those stations that have FlowCapt sensors, the local wind speed measured at this (snow) station can also be used for the index calculation without additional scaling (Eq. (3)), this can be regarded as the most direct verification of the DI. We present and compare the local DI on the one hand and the standard DI, which is based on the wind station measurement, on the other hand with the local FlowCapt reading separately. Fig. 7 shows the DI during the winter 2005/2006 at the IMIS station Davos Frauentobel (2330 m a.s.l.), which is an important station to assess snow drift with respect to a potential avalanche reaching the settlement of Davos Frauenkirch. The comparison in Fig. 7A shows that there is a very good agreement between the snow drift events measured by FlowCapt and the DI. Note that the FlowCapt output is arbitrarily scaled against the DI (cm additional lee slope loading per day) such that the peaks in both time series have similar sizes. This scaling is kept constant for all comparisons presented here. One notable disagreement between FlowCapt and the DI is on the first significant drift event at the end of November 2005. The drift event is

Fig. 6. Photograph of the FlowCapt Drift sensor.

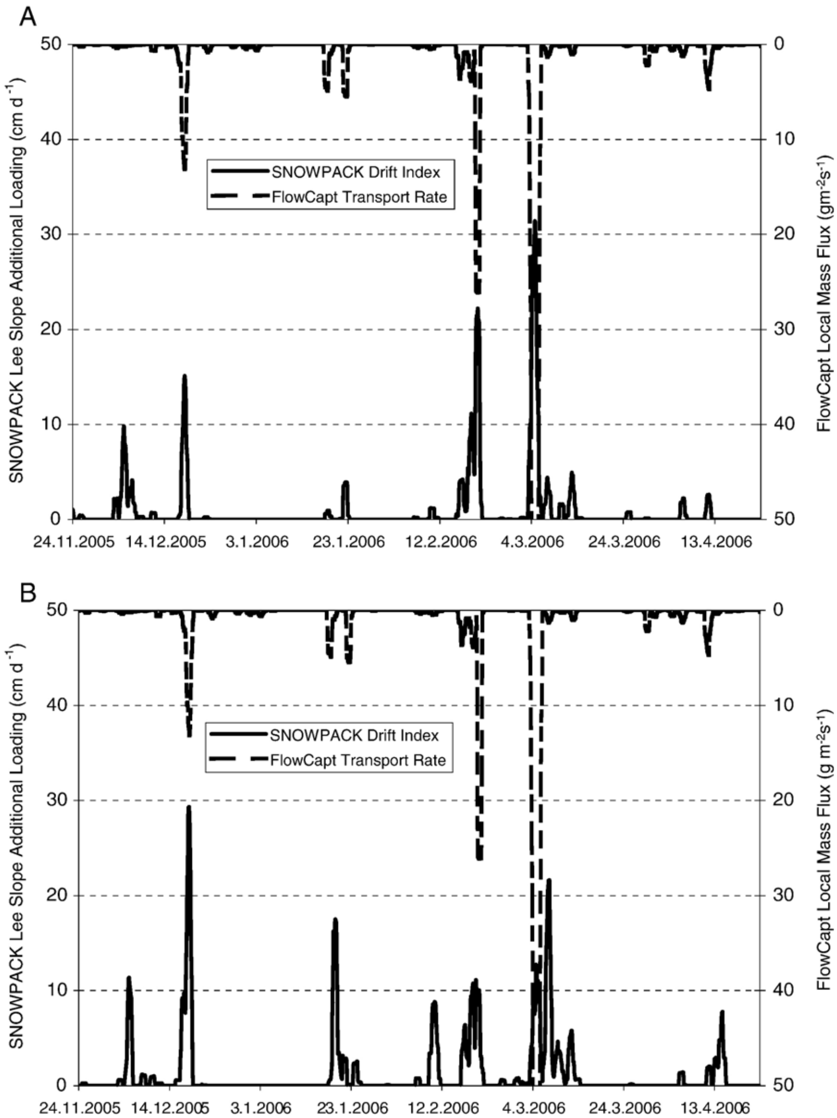

Fig. 7. A: Comparison of the local SNOWPACK drift index with FlowCapt acoustic measurements of local mass transport rate for the IMIS station Davos Frauentobel and the winter 2005/2006. The measured mass transport rates correspond well to the drift index predictions in terms of events and relative magnitude. B: As Fig. 7A but for the regional SNOWPACK drift index using the wind measurements from Krachenhorn and the snow simulations at the Davos Frauentobel IMIS station.

very pronounced in the SNOWPACK DI but insignificant in the FlowCapt reading. Since in this period, snow transport was also reported in the national avalanche bulletins to be a local factor for potential avalanche release (not shown) it may be that the FlowCapt sensor had a problem with riming. Another surprising feature is the huge maximum peak in the FlowCapt reading of 90 g m^{-2} s^{-1}, which is more than 3.5 times the size of the next largest peak of approximately 25 g m^{-2} s^{-1}. Considering the many simplifications behind the SNOWPACK DI and the difficult FlowCapt technology, the general agreement of the two time series is very remarkable, however. The operational DI using the Krachenhorn wind measurement (see Section 4.2) is compared to the same FlowCapt time series in Fig. 7B. The same snow drift periods are detected by the DI as for the local Frauentobel DI but slightly longer average durations are calculated. The magnitude of the DI has changed more considerably. This confirms the finding of the comparison in Section 4.2 that even though over a distance of the order of 10 km the same snow drift periods are found, their magnitude may already change.

Fig. 8 shows the same comparison as Fig. 7 for the station Frasco in the Ticino south of the Swiss Alps. The station is at lower altitude (2200 m a.s.l.) and because of warmer temperatures in the Ticino, less snowdrift is expected. This is confirmed by Fig. 8. While in Davos maximum predicted additional lee slope loading could be more than 30 cm for a single event, it is less than 12 cm in Frasco. This difference is even more pronounced for FlowCapt (90 against 11 g m^{-2} s^{-1}). The local comparison at the FlowCapt IMIS station Frasco Costa shows again that all major snow drift events are present in the FlowCapt readings as well as in the SNOWPACK DI. Here, the SNOWPACK DI appears to underestimate or miss some of the smaller events shown by FlowCapt. It may be that some of the very small FlowCapt readings are caused by wind only, which cannot be judged from the data available. Fig. 8B shows the comparison of the same FlowCapt readings with the IMIS station Frasco Efra (2150 m a.s.l.). Frasco has no separate wind station, thus the SNOWPACK DI uses

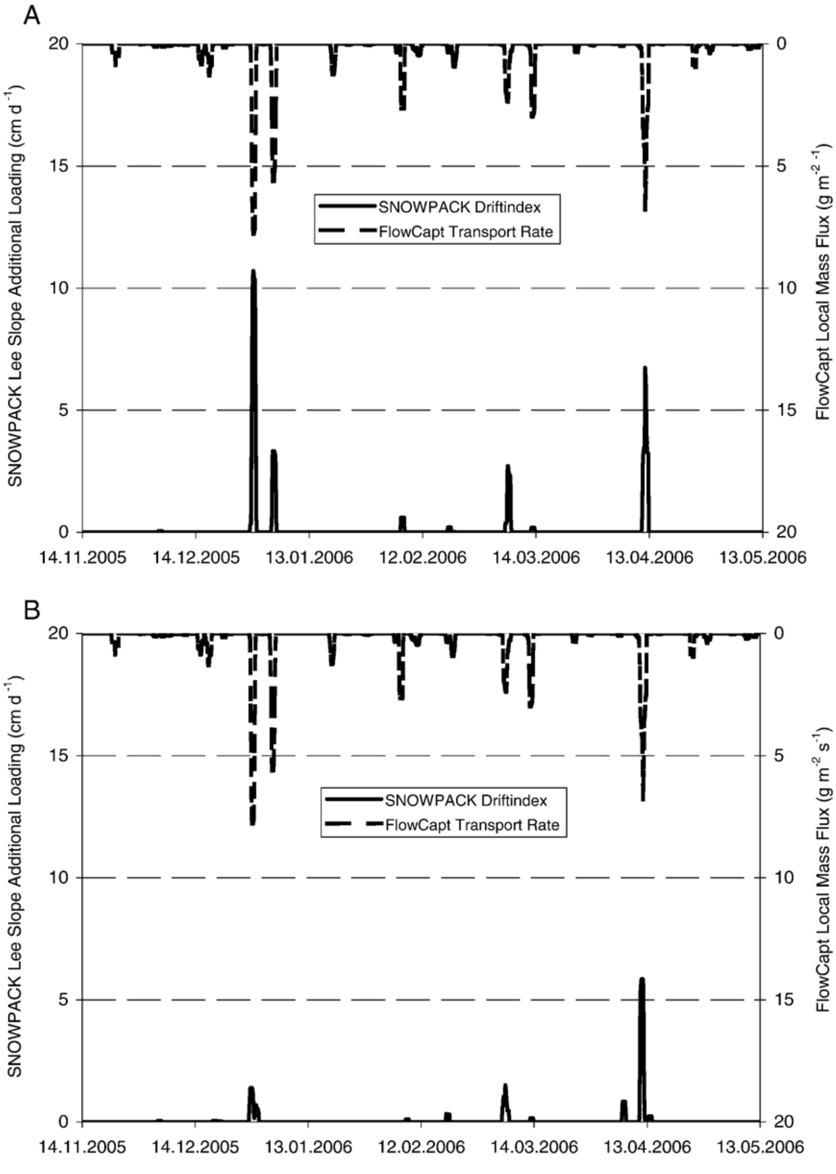

Fig. 8. A: Comparison of the local SNOWPACK drift index with FlowCapt acoustic measurements of local mass transport rate for the IMIS station Frasco Costa during the winter 2005/2006. B: As Fig. 7A but for the local SNOWPACK drift index using the wind measurements and snow simulations from the Frasco Efra IMIS station.

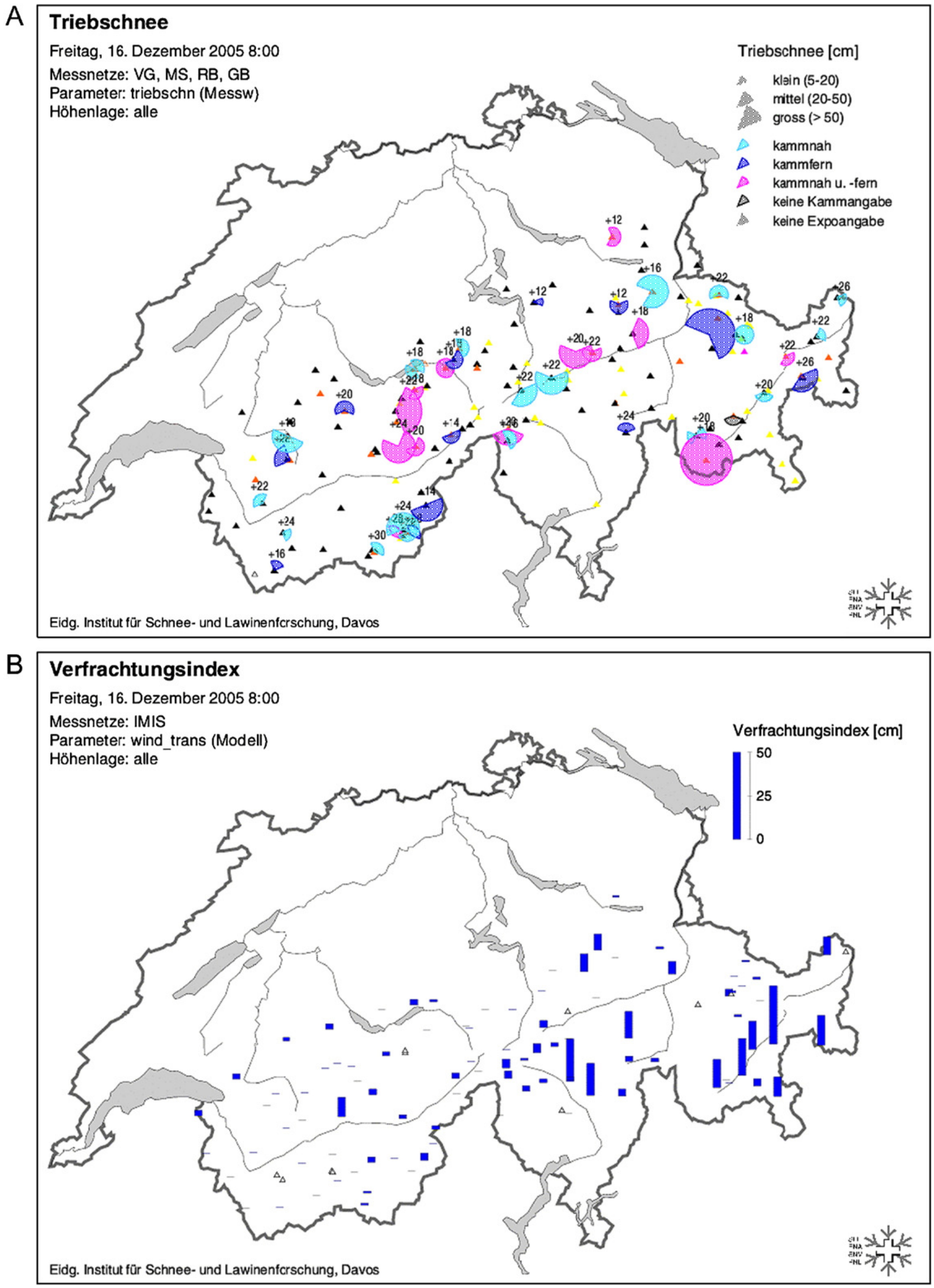

Fig. 9. A: Plot of observed snow drifts over the Swiss Alps on December 16, 2005 as reported by experience observers, B: Plot of simulated drift index over the Swiss Alps on December 16, 2005.

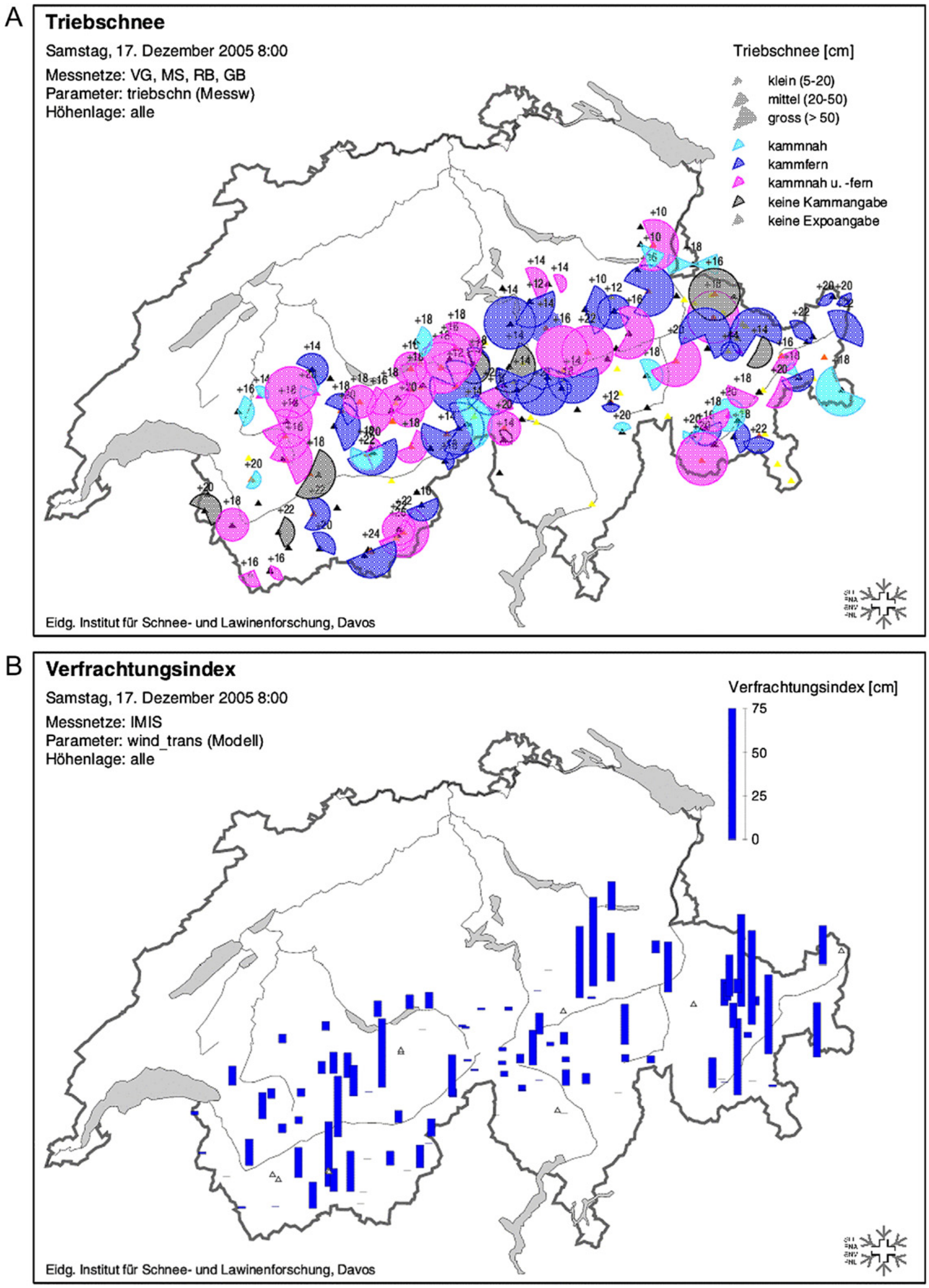

Fig. 10. A: Plot of observed snow drifts over the Swiss Alps on December 17, 2005 as reported by experienced observers, B: Plot of simulated drift index over the Swiss Alps on December 17, 2005.

also the local snow station wind. Apparently, the Flow-Capt sensor at Frasco Costa is at the right location since the DI values at Frasco Efra are even lower. The comparison confirms the findings in the landscape Davos: Snow drift events are regional features but their magnitude may vary locally.

4.4. Comparison of simulated and observed drift patterns over the Swiss Alps

In addition to evaluating the time series of the DI at individual stations, the avalanche warning service is also interested to see which areas of the Swiss Alps are particularly affected by drifting snow. Fig. 9A is a graphical representation of the observed drift reports for all of the Swiss Alps on December 16, 2005. Every circle is from one local observer, who are mostly located in the Alpine valleys. The magnitude of the estimated drift is represented by the radius of the circles and the segments show the aspect of loaded slopes. The two digit number on the plot shows the altitude above which snow drifts are observed, e.g. +18 means that drifts are observed above 1800 m a.s.l. Fig. 9B gives the snow drift pattern as predicted by the SNOWPACK DI, where the DI value (Verfrachtungsindex) is given by the height of the bar. The comparison shows that large DI values are predominantly observed in the eastern part of the Swiss Alps. The SNOWPACK DI predicts somewhat lower drift in the Alps southwest of Berne than would be expected from the observations. In addition to the spatial separation between observers and weather stations, a frequent reason for underestimated DI values is rimed wind speed sensors. In fact, low wind speeds have been measured at the wind stations in the area at the time in question (not shown). Whether the wind speeds were indeed low or whether riming had played a role can not be decided neither from the data nor from the observations available. One day later, on December 17, large DI values are observed (Fig. 10A) and simulated (Fig. 10B) for all of the Swiss Alps.

5. Conclusions and outlook

The estimation of the additional snow mass in lee slopes primarily for the assessment of avalanche danger has been a persistent problem. Thus far, reports by observers have been the most reliable and often the only source of information on blowing and drifting snow. This contribution shows that there are two new methods that have proven to be effective: First, for locations with a standard automatic weather station, snow drift can be estimated using the model drift index DI of the snow cover model SNOWPACK. The new SNOWPACK DI has an improved process representation and uses virtual slopes and aspects to allow erosion in windward and deposition in lee slopes. Simulating snow cover development on those virtual slopes, the snow cover model SNOWPACK provides the framework for DI calculations. The new index can represent local to regional blowing and drifting snow events when compared to observations and distinguishes between regions of high and low DI values. Second, the drift sensor FlowCapt can be used to measure local snow transport. The sensor has proven to work reliably for full winter seasons at carefully selected locations. The high correlation with the SNOWPACK DI, which is constructed using many assumptions and simplifications, appears to validate both methods. From the limited comparison presented here, it can also be said that calculated SNOWPACK drift indices based on local wind measurements at the site of snow simulations have a higher correlation with the locally measured flux than the regional DI, which is based on a wind measurement at some distance. This is not invalidating our approach of trying to publish a regional DI. By contrast, it shows the necessity of having a regional drift assessment, because drifting snow has a high local variability.

In the future, the one-dimensional slope simulations discussed here will also be investigated with respect to stability estimations (Bellaire et al., 2006; Schweizer et al., 2006). Ultimately, the one-dimensional simulations need to be replaced by three-dimensional simulations, which take the snow distribution due to wind transport and the local variability of snow mass and snow characteristics into account. While three-dimensional assessments become now available for research, their operational implementation will still take some time and the one-dimensional assessments presented here will therefore continue to be useful.

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