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Changes in the surface roughness of snow from millimetre to metre scales

S.R. Fassnacht [a,*], M.W. Williams [b], M.V. Corrao [a]

[a] Watershed Science Program, Colorado State University, 400 University Avenue West, Fort Collins, CO 80523-1472, USA
[b] Department of Geography, University of Colorado, Boulder, CO 80309, USA

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ABSTRACT

The roughness of snow influences the movement of air across the snow surface and resulting transfers of energy. Here we focus on the roughness of the snowpack surface to determine its range of variability for different snow conditions (e.g., time since last snowfall), across spatial scales that ranged from 0.01 cm (card) to more than 1000 cm (transect) or more than 5-orders of magnitude, and due to the deposition of aeolian constituents. Digital photogrammetry of snow surfaces was used to compute two roughness metrics at two mountain sites in north-central Colorado. These metrics are the random roughness (*RR*) that disregards the spatial structure and the fractal dimension (*D*) computed from variogram analysis.

At the crystal scale, *D* is between 1.67 (card) and 1.60 (board), which increases to 1.77 between 0.1 and 1.0 cm. At longer scales, *D* is 1.53 (board) to 1.56 (transect). There was no significant change in surface roughness during the accumulation season, with *RR* values at about 0.002. During the melt season the surface roughness doubled, with the *RR* values increasing from about 0.002 to 0.004. Snow was more rough parallel to the wind when dunes were present, and roughness varied spatially. The average *RR* value computed for the white snow surface of 0.014 is substantially greater than the value computed for the red dust surface of 0.0032. Due to undulations of smaller amplitude and as a result of the dust itself, the red dust surface is more random (*D* is 2.62 versus 2.23 for the white snow). Our results show that there is consistency in roughness over different scales, yet large scale processes (e.g., wind and radiation activity) influence the magnitude of roughness metrics much more than small scale processes (e.g., crystal form and metamorphism).

1. Introduction

Snow surface roughness is a measure of the variability of surface microtopographic features or roughness elements at different scales. Fine scale features exist at the millimetre to centimetre size due to variation in grain structure, while large scale features develop due to meteorology (especially wind), topography, and underlying vegetation (Munro, 1989; Smeets et al., 1999). The large-size features include sustrugi and ice hummocks. Knowledge of the surface roughness of snow fields, glaciers, and ice sheets is important for a variety of reasons, including understanding surface–atmosphere interactions and how snow is remotely sensed in both the visible and microwave portions of the spectrum. Destructive metamorphism of freshly fallen snow is rapid (McClung and Schaerer, 2006) and can dramatically change the roughness of a fresh snow surface.

The nature of a snowpack surface is influenced by its meteorological history while subsequently providing feedback to the near surface meteorological conditions. For the movement of air across the snow surface, the aerodynamic roughness length (z_o) is a parameter describing the surface roughness that is used to determine sensible and latent heat fluxes between the surface and the atmosphere. Estimating values of z_o is difficult using visual or physical measurements to characterize surface roughness elements (Lettau, 1969) and a review by Andreas (2002) has highlighted the complexity in aerodynamic turbulence across a snow or ice surface.

Several metrics have been used to describe the surface roughness of various media. Huang (1998) provides a summary of the key metrics used in soil science. Indices developed for soil science include the random roughness (Kuipers, 1957), the sum of the absolute slopes between various distance intervals (Currence and Lovely, 1970), the product of the microrelief index (mean absolute deviation of elevation from a reference plane) and the peak frequency (number of elevation peaks per unit transect length) (Romkens and Wang, 1987). Fassnacht et al. (2009) used these roughness indices to determine the roughness of a snowpack

* Corresponding author. Tel.: +1 970 491 5454; fax: +1 970 491 6754.
E-mail address: srf@cnr.colostate.edu (S.R. Fassnacht).

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surface. These indices were compared to several roughness measures which quantify the spatial structure of the surface, including the semi-variance (Brown, 1987), autocorrelation (Huang, 1998), and power spectral density (Currence and Lovely, 1970). The semi-variance is a geostatistical measure used to characterize the spatial variability of an attribute and to compute the range over which data are correlated. A variogram is a plot of the semi-variance versus the lag distance or average distance between measurements. A power–law slope can be used to compute the fractal dimension (D), equal to the dimension of space plus one minus one-half of the exponent in the power relationship (Huang, 1998). A change in the power–law variogram defines a scale break (SB); often the semi-variance becomes constant (D approaches 2 for a profile) at lag distances greater than the SB, implying randomness.

This paper focuses on measurements of the surface roughness of seasonal snowpacks to determine its range of variability for different snow conditions (e.g., time since last snowfall) and across spatial scales over five orders of magnitude. Digital photogrammetry and physical measurements were used to characterize the surface roughness. Stereo-photogrammetry has been used by various researchers to estimate soil roughness indices (e.g., Taconet and Ciarletti, 2007). However, since a snow surface can be penetrated by a board with minimal disturbance (e.g., Cline et al., 2003), digital photogrammetry of the snow surface can be used to estimate roughness without the requirement of stereo imagery. Two metrics to evaluate surface roughness were used: a roughness index called random roughness (RR) and the fractal dimension (D), following the protocols developed by Deems et al. (2006) for analyzing the spatial variation of snow depth.

The near-surface snow environment was sampled to address the following objectives: (i) to determine how the roughness of the snow surface may vary with changes in spatial scale, (ii) to determine how the snow roughness changes with time at a point, and (iii) to evaluate how atmospheric impurities may influence surface roughness. Scaling in surface roughness was assessed from the crystal scale at an approximate resolution of 0.05 mm to the multi-meter scale at a resolution of 0.25 m. How surface roughness changes with time at a point was addressed using 1 m squares of roughness measurements, and included evaluations of potential anisotropy. The last objective was addressed by comparing a "red snow" surface from a snowpack that contained aeolian dust with an adjacent surface of "clean snow" without red dust.

2. Material studied, area descriptions, methods, techniques

2.1. Location

Measurements were collected at two different sites in north-central Colorado. The first two objectives were addressed with data collected during the winter of 2006 (Fig. 1) and on December 19th, 2007 at a site called the Michigan River Meadow (Fig. 2). The winds blow predominantly from the west (Fig. 2) with a fetch of 300 m across this flat and open site. Maximum wind velocities approach 10 m s^{-1} and enable the transport of snow particles which can create dunes. Snow covers the underlying 0.3–0.4 m tall shrubs by early December and these shrubs remain snowcovered until mid-May. The Michigan River Meadow site is located 100 m north of Colorado Highway 14 approximately 3 km west of Cameron Pass (at latitude 40°31′N and longitude 105°55′W). This site is near the location of the Natural Resources Conservation Service (NRCS) Joe Wright snow telemetry (SNOTEL) station. This site receives approximately 1100 mm of precipitation annually with an average peak snow water equivalent of 680 mm (Ewing and Fassnacht, 2007).

Sampling for the third objective occurred due west of the city of Boulder, at the Niwot Ridge Long-Term Ecological Research

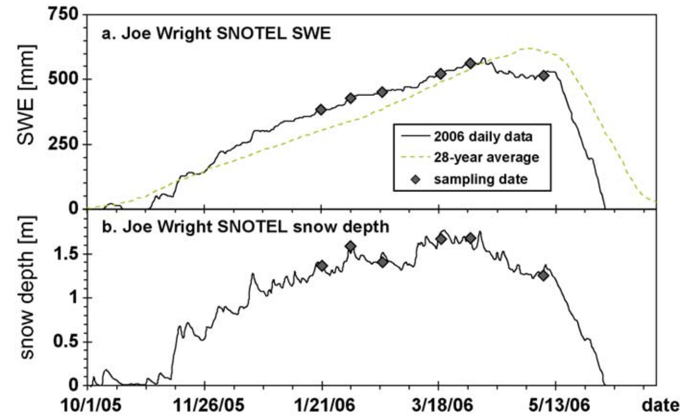

Fig. 1. Time series of SNOTEL (a) snow water equivalent (SWE) and (b) snow depth at the Joe Wright SNOTEL station (near the Michigan River Meadow study site) for the 2005–2006 snow season. The 28-year average SWE and the six sampling dates are presented.

Fig. 2. Photograph of the Michigan River Meadow site looking west or into the predominant wind.

program. The site receives approximately 1000 mm of precipitation annually, of which a majority falls as snow (Williams et al., 1996). On February 14th and 15th, 2006, a blowing dust event deposited red dust across most of the state of Colorado (Losleben et al., 2006). The source of this dust was likely the Colorado Plateau to the southwest of the study site (Painter et al., 2007), but it may have Asian origins. On the ridge above the Soddie Laboratory (at latitude 40°3′N and longitude 105°33′W) at an elevation of approximately 3380 m, the red dust was not distributed uniformly because of differences in wind patterns, providing areas covered with red dust adjacent to control "clean" snow areas with no red dust. The red dust at and near the snow surface was then covered by subsequent snowfalls. Sampling occurred on June 13th, 2006 on two areas with almost identical topographical characteristics (Fig. 3), i.e., they were on the same hill, had the same slope and aspect, and separated by a distance of only 10 m, so any differences in surface roughness were likely the result of the impurities on and near the snow surface. The site was visited about weekly throughout the snow accumulation and melt seasons and no differences in surface roughness were noticed until the red snow surface was exposed during snow melt. The predominant wind direction was from west to east with an open fetch in excess of 500 m.

2.2. Data collection

To estimate snowpack roughness using digital photogrammetry, a black board (0.79 m or 1 m) or crystal card (0.097 m) was used. The board was inserted vertically into the snowpack so as to

Fig. 3. Photograph of the Niwot Ridge site looking at the lower dust covered surface and upper 'clean' no dust area. The predominant wind direction is from west to east or left to right.

yield a continuous interface between the snow and the board. For some of the measurements later in the snow season, such as those in the clean snow at Niwot Ridge, a rubber mallet was required to drive the board into the snow to yield a continuous interface. Digital photographs were taken of the snow board (or crystal card for a finer resolution) and the interface with the snow surface using a SONY© DSC 707 with an image size of 2560 × 1920 pixels. The crystal card has a spatial extent of 9.7 cm at a resolution of 0.043 mm. The snow board has a spatial extent of 79 cm at a resolution of 0.32 mm. The snow depth transect has a spatial extent of 25 m at a resolution of 0.25 m. This sampling protocol provides spatial information that ranges from 0.01 cm to more than 1000 cm, or more than 5-orders of magnitude.

For the spatial variation analysis, the snow board was inserted into the snow at 10 cm intervals to yield a 1 m segment of 11 parallel roughness images. The board was rotated 90° and inserted 11 more times at 10 cm intervals over an adjacent undisturbed snowpack to produce a second set of 11 parallel roughness images.

The snow roughness data for the topographical scale were derived from overlaying snow depth data on an interpolated ground surface. The snow depth data were recorded to the nearest 0.01 m at 0.25 m intervals along a 25 m transect with a 0.015 m diameter depth probe. Prior to initial snow accumulation, a total station was used to survey the ground surface along and beside this transect. These data were sampled at any elevation change or undulation greater than 0.05 m yielding an average increment of 0.35 m between points. These data were then interpolated using the inverse distance weighted method. This interpolated surface was at most 0.05 m different than one derived from kriging. This implies that the maximum interpolation error is approximately 0.05 m.

2.3. Image analysis

Once the photographs were taken, the images were transferred to a computer. Images were clipped so that the snowpack–snow board interface at the left and right of the image became the boundary of each new image. The image was converted to black and white and was saved as a target image file format (TIFF). The ESRI ARC/INFO© software was used to convert the TIFF format into a raster or grid file in ASCII text file format. The raster file was converted into a text file of *X*, *Y*, *Z* coordinates. For a single board, the *Y* values were set to zero. For multiple boards, the coordinates of each end of the board were used to put the board into three-dimensional space. A digital number of 128 was used as the threshold between snow and no-snow (Fassnacht et al., 2009).

2.4. Data analysis

Two metrics to evaluate surface roughness were used: a roughness index called random roughness (*RR*) and the fractal dimension (*D*). The *RR* index was computed as the standard deviation of the elevations from a mean surface. The *D* measure was determined from linear segments manually examined on variograms with log–log scales. The variograms were generated for each data set representing the snow board interfaces using all data pairs (points in *X*, *Y*, *Z*) organized by their distance apart.

3. Results and discussion

3.1. Scaling of roughness

The snow surfaces at three scales on December 17th, 2007 are presented in Fig. 2. The snow surface at the coarse-resolution transect scale (Fig. 4a) visually appears to be smoother than at board scale (Fig. 4b) and at the crystal scale (Fig. 4c). However, the *RR* of the surface decreases by 1.5 orders of magnitude as scale decreases from the transect to board scale, from 0.063 to 0.0014. The *RR* also decreases going from the board to crystal scale, but only by a factor of two. Thus, *RR* values indicate that the surface roughness decreases by almost two orders of magnitude as the measurement spacing (resolution) increases while the extent of the measurements decreases by about 3 orders of magnitude.

Variogram analyses show that the semi-variance varies widely with scale (Fig. 5). At the scale of the crystal card, there is a sill at about 0.005 cm^2. At the scale of the snow boards the sill increases about an order of magnitude to 0.05 cm^2, and at the transect scale the sill is approximately 50 cm^2. Consistent with the *RR* estimates, it appears that as the measurement spacing becomes coarser, the semi-variance of the snow surface increases.

Three sets of snow surface fractal dimensions are present in Fig. 5. At the crystal scale, *D* is between 1.67 (card) and 1.60 (board), which increases to 1.77 between 0.1 and 1.0 cm. At longer scales, *D* is 1.53 (board) to 1.56 (transect). When the concurrent snow depth and ground surface are combined to yield the snow surface in the transect dataset (Fig. 6), the snow surface (D = 1.56) follows the snow depth (D = 1.64) when measurements are up to 5 m apart, and the ground surface (D = 1.11) at longer distances.

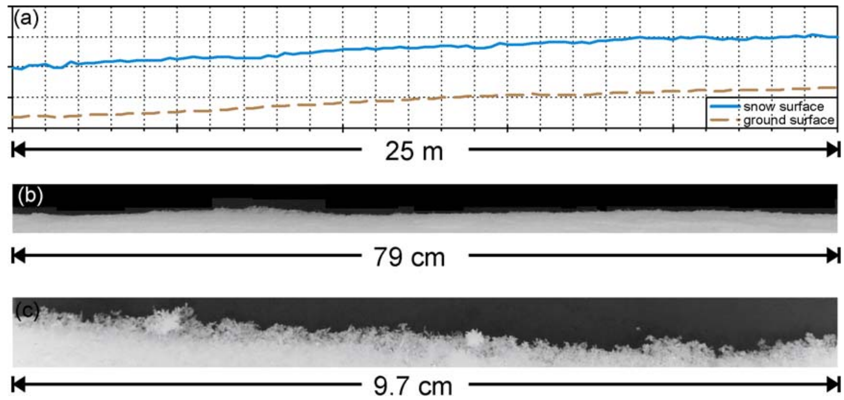

Fig. 4. Proportionally scaled representations of the snow surface on December 17th, 2007 in the meadow by the Joe Wright SNOTEL station for (a) the 25 m snow depth transect at a 1 m resolution, (b) the 79 cm snow board at a 0.32 mm resolution, and (c) the 9.7 cm crystal card at a 0.043 mm resolution. The snow depth transect was derived from snow depth measurements overlain on a ground surface map interpolated from measurements at a 0.1 m interval. The snow board and crystal card are black and white 8-bit photographs of the snow surface interface.

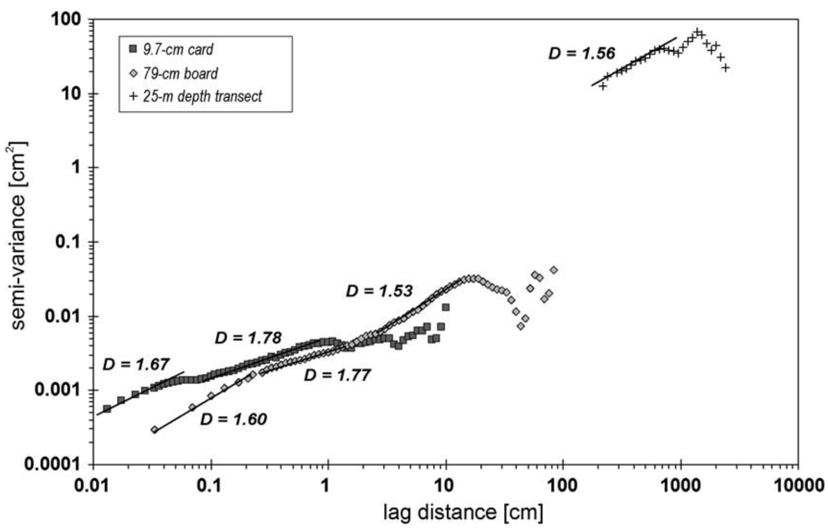

Fig. 5. Snow surface roughness across four orders of magnitude of resolution and two-and-a-half orders of magnitude of extent. The crystal card covers 9.7 cm at a 0.043 mm resolution, the board covers 79 cm at a 0.32 mm resolution, and the depth transect covers 25 m at a 0.25 m resolution.

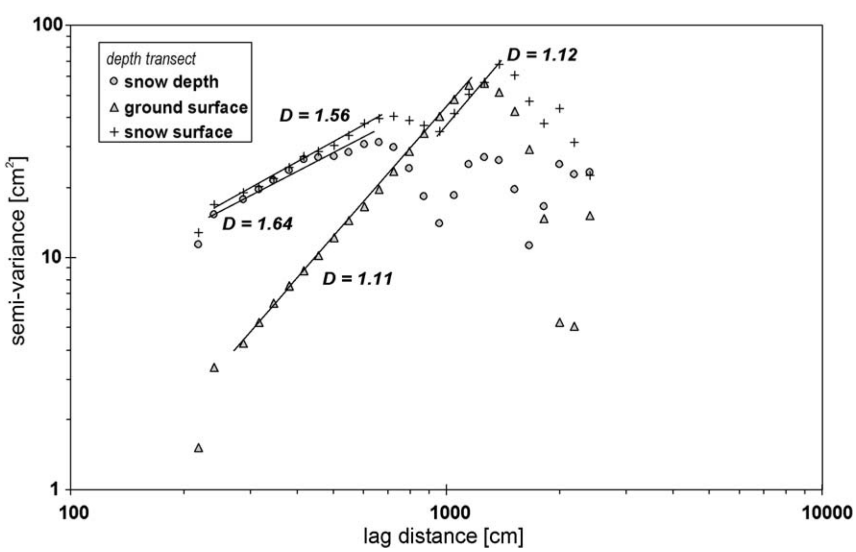

Fig. 6. Snow depth, ground surface, and snow surface semi-variograms for the 25 m depth transect. The snow depth was measured and added to the ground surface interpolated from 0.1 m measurements to yield the snow surface.

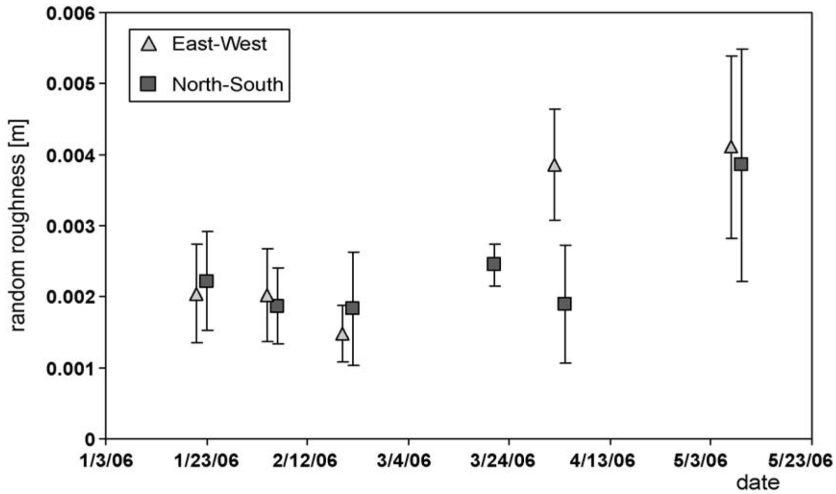

Fig. 7. Temporal change in random roughness for the 2006 sampling dates at the Michigan River Meadow site averaged for the east–west and north–south oriented boards, with error bars of one standard deviation.

While care was taken to ensure that the snow depth transect matched the ground surface, future work should concurrently measure the ground elevation and snow surface elevation.

Deems et al. (2006) computed the fractal dimension of approximately 2.47 with a scale break between 15 and 41 m for a snow depth surface estimated from airborne LIDAR at three 1 km^2 sites in northern Colorado. The rough surface roughness presented herein is for a plane, thus, if the difference between the surface and plane fractal dimensions is equal to one, both features have the same fractal characteristic. It should be noted that due to the scalable nature of fractal dimensions, it is appropriate to compare the fraction of dimensions for different space, i.e., a line/curve versus a surface. These snow depth D values are very similar to the surface roughness D values for the board (0.03–0.15 m lag distance) and the transect (2–8 m lag distance). The LIDAR data were at a horizontal spacing of 1.5 m, while the board were at a spacing of 0.4 mm (3.5 orders of magnitude finer) and the transect at a spacing of 0.25 m (<1 orders of magnitude finer). Blöschl (1999) showed that D was the same across multiple scales for datasets representing similar snow properties, in particular, snow covered area. The difference in D between the Deems et al. (2006) snow depth data and the surface roughness presented in this paper may be because the data represent inherently different properties—snow depth versus snow surface measurements.

3.2. Temporal variation in surface roughness

Snowfall and accumulation patterns in 2006 reflected average conditions for the 28-year history of the Joe Wright SNOTEL station. Snow accumulation began in early November and reached maximum accumulation on April 7th. The peak snow accumulation of 608 mm was similar to the long-term average of 680 mm (Fig. 1). The snow pack became isothermal at 0 °C on approximately April 18th, with significant snow melt starting on May 10th.

The surface roughness was measured at the same site on 6 sampling dates (Fig. 1) using the board (at an approximate 0.4 mm resolution). The sampling dates ranged from January 12th to May 7th, spanning the snow accumulation season (samples collected from January 12th to February 19th), maximum snow accumulation (samples collected on March 19th and April 2nd), and into the melt season (samples collected on May 7th).

Since heavy snow was falling during the March 19th sampling, no east–west data were collected. Fresh snow had fallen within 24 h of sampling on January 12th and February 4th. Snow fell two days prior to the April 2nd sampling, and four days prior to the February 19th sampling. Snow fell 12 days prior to the May 7th sampling with 2 rain events occurring in between. While the top layer of the snowpack was less dense when the time between the last snowfall and sampling was shorter, no correlation was found between roughness characteristics and days since the last snowfall event observed at the SNOTEL station.

The *RR* values were consistent at about 0.002 for the five samples collected during the snow accumulation season and at maximum accumulation (Fig. 7). During the melt season the surface roughness doubled, with the *RR* values increasing from about 0.002 to 0.004. It is expected that roughness increases during melt due to localized melt patterns and no snowfall creating a new smooth surface. Five of the six sampling dates showed no significant differences in *RR* oriented along north–south and east–west axes, with $p > 0.05$. These results suggest that directional differences in the surface roughness of snow are limited when comparing the surface parallel to the predominant wind direction versus perpendicular to it.

Geostatistical analysis of the data collected on April 2nd shows that there is twice the *RR* in the east west direction compared to the north south direction (Fig. 7), illustrating the anisotropy at this time. Variograms for north south and the east west directions are shown in Fig. 8a and b, respectively. The surface roughness varied over space; the north–south oriented boards (Fig. 8a) placed perpendicular to the wind were more consistent than those parallel (Fig. 8b) to the wind. The semi-variances of boards placed in the same direction were not always the same and could even present different scale breaks (Fig. 8b).

As the snow season progressed, the fractal dimension decreased for the short-scale (Fig. 9a) while it increased at a lesser rate for the large scale (Fig. 9b). The shorter scale D varied from 1.75 to 1.4 while D varied from 1.3 to 1.55 at the longer scale. The directional differences in D were not consistent at either scale or with the *RR*. The *RR* values are dominated by larger scale surface variations over smaller scale ones due to their magnitude, thus the *RR* values reflects the longer scale D (Figs. 7 and 9). As the smaller scale D decreases, both the *RR* and longer scale D increase. Since the *RR* values are computed from standard deviation of the detrended surface, variability at the longer scale tends to be larger thus having a more substantial influence on *RR*.

There are different scales of variability, from crystal scale to wind deposition scale (seen directionally on April 2nd, 2006 in Figs.

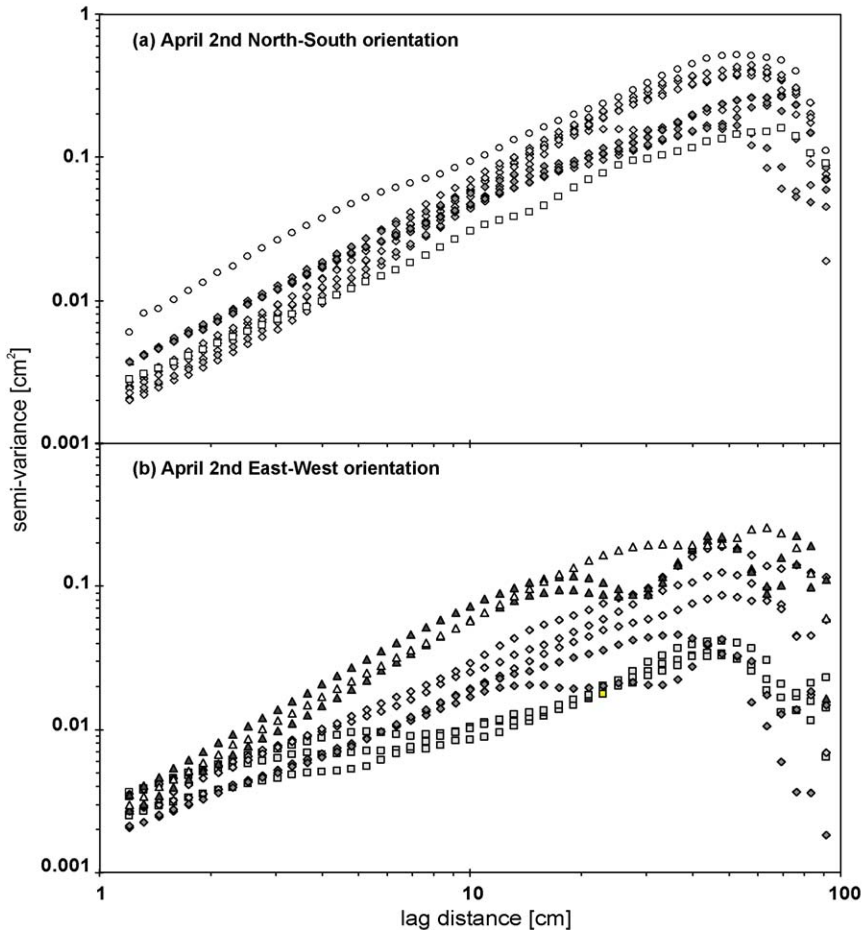

Fig. 8. Variogram for the 11 individual boards each at 10 cm apart on April 2nd, 2006 at the Michigan River Meadow site in the (a) north–south orientation, and (b) east–west orientation. Each sequence of symbols represents an individual board.

7–9b), to topographic scale. Deems et al. (2006) showed the transition from vegetation influences to topographic influences. Those data were at a 1.5 m resolution and covered 1 km^2, so this transition is not seen in this work due to the smaller extent. However, larger scale snow processes influences the surface more than small scale snow processes. For example, local topographic features present in a ground surface influence the snow surface roughness during accumulation which interacts with wind induced surface changes (Fig. 6). Such features supercede crystal scale influences (Fig. 5).

The crystal scale roughness could not be determined using the boards. Imagery at the spacing (resolution) of the card or a digital camera at an order of magnitude higher resolution (>50 mega-megapixels) is needed to acquire images necessary to determine crystal scale roughness features. The card is needed to acquire images at the spacing (resolution) necessary to determine crystal scale roughness features. Information from the crystal card therefore could not be used for the temporal analysis, so the influence of different snow crystal forms over time was not investigated. Future work should consider sampling at several spatial resolutions.

3.3. Influences of aeolian dust

The deposition of dust on the snowpack creates a layer with a lower albedo than snow without the dust layer (Painter et al., 2007). This difference can immediately influence the properties of the snowpack, and will influence snowmelt characteristics, especially when re-exposed. The aeolian dust event with the distinctive red colour was deposited four months prior to the sampling of the snow surface roughness (Losleben et al., 2006). From a SNOTEL site, 300 m below this study site, 90 mm of SWE accumulated after the aeolian deposition in at least four snowfall events, completely covering the dust. The snow surfaces at clean and red dust locations were very similar prior to deposition. More snow fell at the study site during the four months between deposition and sampling than at the lower elevation SNOTEL site (NWTLTER); the snow was about 2 m deep at the time of surface roughness sampling, while it had all ablated a month earlier at the SNOTEL site. Painter et al. (2007) showed that dust exposed during melt will remain at the snow surface, illustrating that the differences in roughness due to dust that we observed occurred during melt and not during the snow accumulation season.

Visual inspection of the snow surface showed large differences in surface properties between "clean" and "red" snow (Fig. 10). These photographs (Fig. 10) were representative of the clean and red snow surfaces (Fig. 11). The clean snow surface was white in colour, the red snow was characterized by a consistent reddish colour. The clean snow surfaces without the red dust (Figs. 10a and 11a) have larger amplitudes of undulation than that of the red dust snow (Figs. 10b and 11b). However, the overall period of the roughness undulation is larger for the red dust snow (0.7–0.8 m) than for the white snow (0.2–0.55 m with an average of 0.38 m) while the variation between individual snow surfaces is larger. The average *RR* of 0.014 for the clean snow was 4.2 times larger than

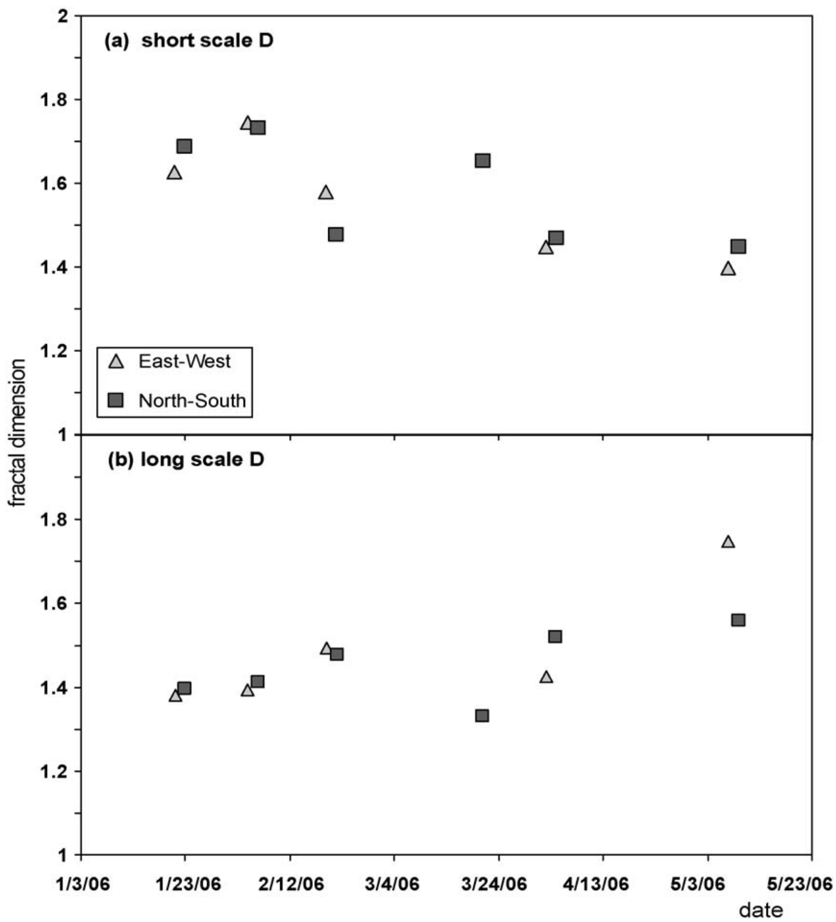

Fig. 9. Temporal change in the average fractal dimension for directional blocks from the variogram analysis for the 2006 sampling dates at the Michigan River Meadow site at (a) the short scale and (b) the long scale.

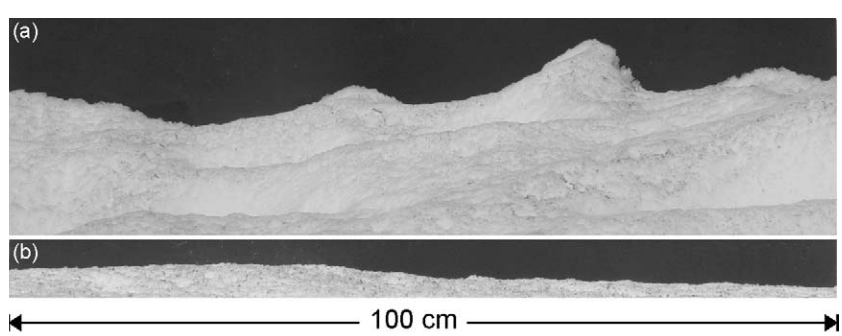

Fig. 10. Sample photographs of (a) the clean snow surface without dust, and (b) the snow surface with red dust taken on June 12th, 2006 on the ridge above the Soddie Laboratory at Niwot Ridge.

the value of 0.0032 for the red dust snow (Fig. 11a and b). The range of variation for the 11 clean snow surfaces (RR = 0.0029–0.0036) was 16 times less than for the 11 red dust surfaces (RR = 0.0089–0.020).

The semi-variance computed for the white snow surface is substantially greater than the value computed for the red dust surface (Fig. 12). Due to undulations of smaller amplitude and as a result of the dust itself, the red dust surface is more random (D is 2.62 versus 2.23 for the white snow). The first scale break occurs at similar distance (0.13 and 0.15 m), while the second scale break is smaller for the white snow (0.28 m versus 0.55 m). The fractal dimension of the second segment (after the first scale break) is larger for the white snow (D is 2.73) than for the red dust snow (D is 2.39).

The second scale break is smaller for the white snow (0.28 m) than the red dust snow (0.55 m) and may be indicative of the periodicity of the snowpack surfaces (~0.38 m for the clean snow and ~0.75 m for the red dust snow). For the clean snow, the larger D above the first scale break indicates a more random surface; this is inverse for the red dust snow as grain scale (and larger) variations are not seen at longer distances. At the sub-grain scale, different fractal dimensions may exist, but this investigation requires data at a finer resolution (e.g., Fassnacht and Deems, 2006).

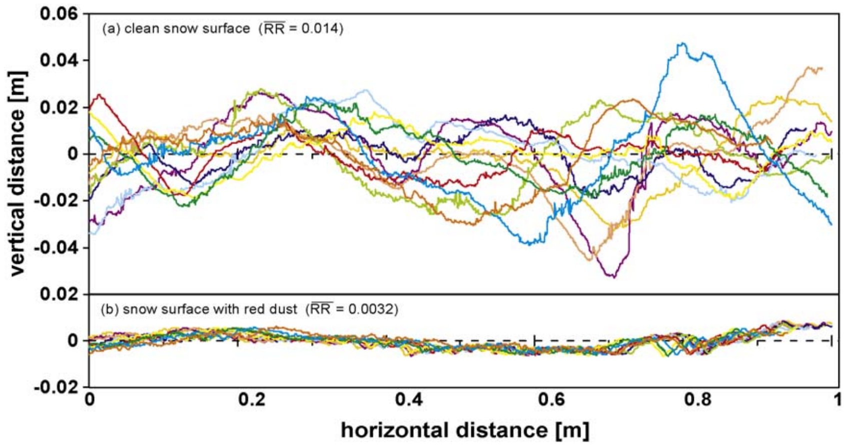

Fig. 11. Surface characteristics of 11 individual boards for (a) the clean snow surface without dust, and (b) the snow surface with red dust. The average random roughness (*RR*) for the clean snow is 0.014 with a range 0.0089 to 0.020, and for the red snow is 0.0032 with a range from 0.0029 to 0.0036.

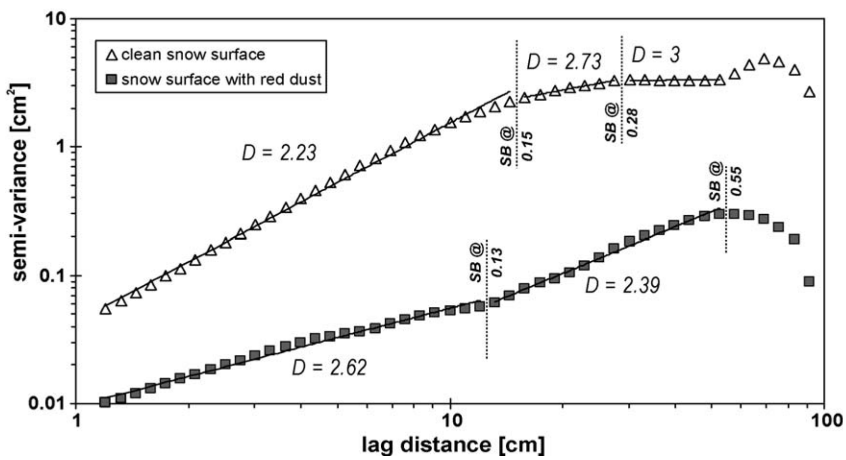

Fig. 12. Semi-variogram for the adjacent clean snow surface and snow surface containing red dust.

4. Conclusion

A roughness index (the random roughness) commonly used in soil science and variogram fractal analysis (yields spatial structure information) were used to illustrate that the snow surface roughness varies. The magnitude of the random roughness and the short-scale fractal dimension (derived from variogram analysis) are inversely proportional.

The relative elevation of the snow surface was measured over four orders of magnitude and shown to be consistent. Large scale processes have a greater influence on measures of roughness as compared to small scale processes. At a fractal dimension between 1.53 and 1.60, snow surface roughness was slightly more random (*D* increases) than snow depth, as presented by Deems et al. (2006).

During snow accumulation, larger scale roughness (greater than crystal size) was consistent, while smaller scale roughness varied over time. Roughness increased while becoming less random (*D* decreases) during snowmelt, as compared to during snow accumulation. Snow was more rough parallel to the wind when dunes were present, and roughness varied spatially. Directional difference between the random roughness and fractal dimension were not consistent.

Wind-blown dust that is deposited during snow accumulation will come to the snowpack surface during snowmelt. These aeolian constituents were shown to decrease the magnitude of the roughness relative to adjacent "clean" surfaces.

The changes in snow roughness at different scales, over time and due to deposited constituents illustrate that the roughness of the snowpack surface is not constant. The magnitude of the surface changes shows the complexity of aerodynamic turbulence across a snow surface. These changes must be estimated to understand the movement of air over snow and the exchange of turbulent fluxes.

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