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Relating variation of dust on snow to bare soil dynamics in the western United States

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Abstract

The deposition of desert dust to mountain snow directly impacts the hydrologic cycle and water resource management through the depression of snow albedo and acceleration of snowmelt. However, the key processes that control the variation of dust deposition to snow are poorly understood. Here we relate the bare soil exposure from the moderate resolution imaging spectroradiometer (MODIS) reflectance data for the period of 2002–2011, with dust loading in snow at downwind mountain sites in southern Colorado, the United States. We found that, for many pixels, remotely sensed fraction of bare soil in the dust-emitting area is significantly correlated with end-of-season dust concentrations in snow, and that the highest number of significantly correlated pixels in the dust-source area corresponds well with the period of peak dust deposition in the mountain snow (April–May). This analysis indicates that surface conditions in the dust-source area may provide first-order controls on emission of dust and deposition of that dust to the mountain snowcover. A preliminary analysis of precipitation records indicates that bare ground cover is strongly affected by prior rainfall in the months preceding the dust-emission season.

Keywords: remote sensing, snow hydrology, desert dust, water resource management, vegetation dynamics

1. Introduction

An increasing number of observations worldwide suggest that light-absorbing impurities strongly influence the energy budget of snow and ice, resulting in increased melting (e.g., Flanner *et al* 2009, Painter *et al* 2012a). There is further evidence that deposition of desert dust and black carbon—the major light-absorbing impurities—to mountain ranges including the Rocky Mountains, Hindu Kush, and

Himalaya, is increasing (e.g., Kaspari *et al* 2009, Kaspari *et al* 2011, Neff *et al* 2008, Ramanathan *et al* 2007, Thompson *et al* 2000, Gautam *et al* 2013). The deposition of desert dust to mountain snow directly impacts snowmelt rates by decreasing snow albedo with influences on basin-scale runoff (Painter *et al* 2007, Steltzer *et al* 2009, Painter *et al* 2010) and regional surface temperature (Cohen and Rind 1991).

The southern Rocky Mountains of the western United States receive multiple dust-deposition events annually peaking during the period from February to May (Painter *et al* 2012b). In this region, radiative forcing by dust in snow is estimated to shorten the snow cover duration by three to seven weeks (Painter *et al* 2007, Skiles *et al* 2012). In the Hindu Kush, desert dust can result in radiative forcing



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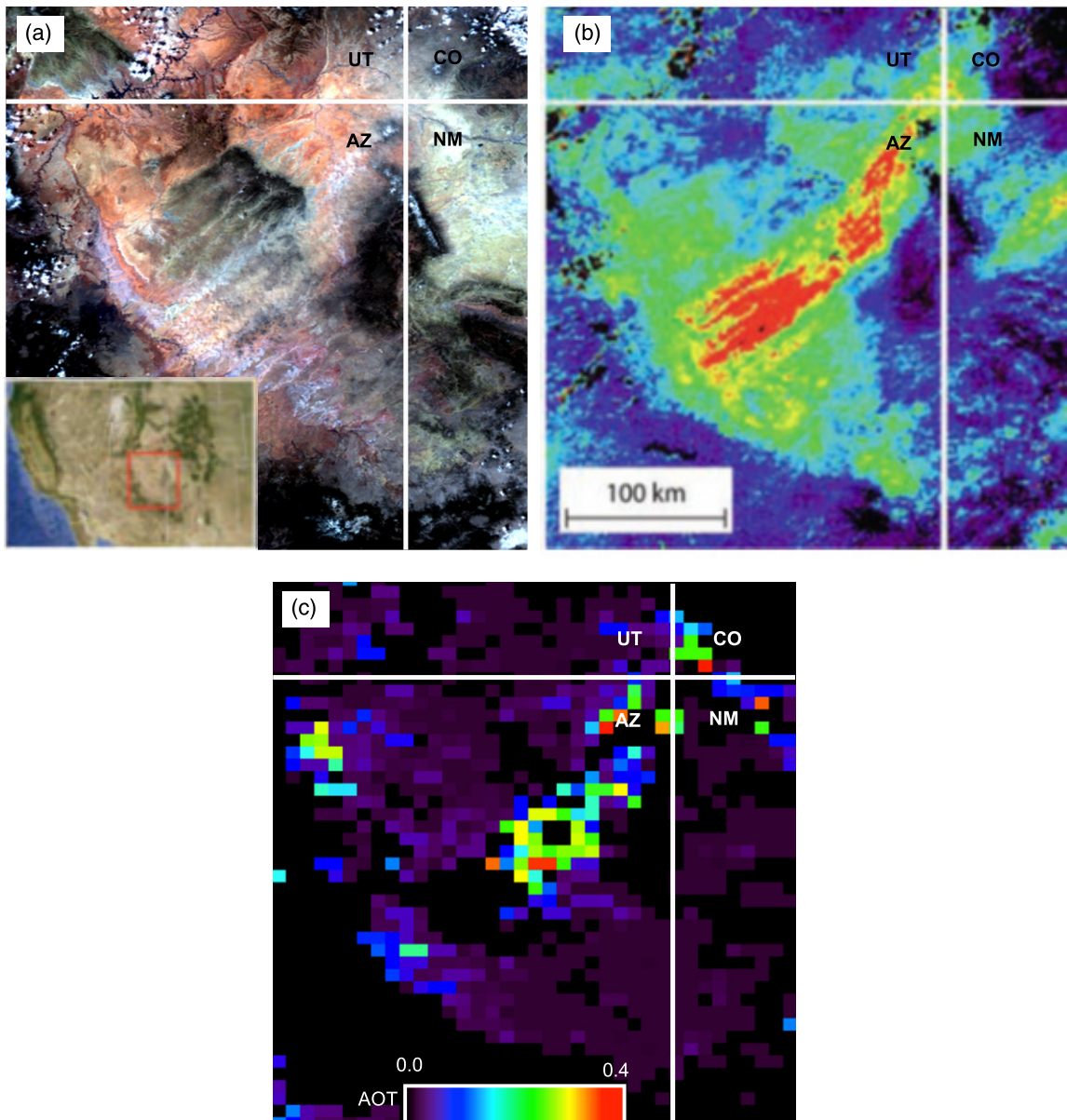


Figure 1. MODIS imagery showing dust plumes in the northeastern Arizona, United States, captured on 8 April, 2009. (a) Real-color composite from MODIS-Aqua Level1B top-of-atmosphere reflectance. (b) Thermal anomaly calculated as the difference between MODIS-Aqua land surface temperature on the day of the event and composited MODIS land surface temperature from the compositing period than contains the event. Areas of high of thermal difference indicate dust plume in the air, with green through red color. (c) MODIS-Aqua Deep Blue 550 nm aerosol optical thickness (AOT) for the same event, with dust plumes indicated by high values of AOT.

reaching 200 W m^{-2} at the surface at the end of the snow-covered season (Painter *et al* 2012b). Such perturbations to the hydrologic cycle have substantial implications for water management, directly impacting the timing and magnitude of water supplies, hydroelectric power, agriculture, and forest fires (Westerling *et al* 2006, Belnap *et al* 2011). This is particularly true when dust-covered mountain snowpack is the source of water for major waterways on which human populations rely (e.g., the Brahmaputra and Colorado Rivers).

The deposition of dust to distal snowcover is variable from year to year and may be related to many processes, including dust emissions in the source regions, dust transport pathways, meteorology in the sources and sinks regions

associated with large-scale atmospheric circulations, and the presence and magnitude of snowfall/rain scavenging (Wake and Mayewski 1994, Prospero and Lamb 2003). One of the ideal locations to address the question of what controls deposition of dust on snow is located in the dust-emission area of Four Corners (where Colorado, Utah, Arizona, and New Mexico touch) and the nearby dust-deposition area in the San Juan Mountains in the southwestern Colorado. Large-scale dust storms have been frequently recorded in remote sensing imagery over four corners and northeastern Arizona in the western United States (figures 1(a) and (b)). The provenance of dust in the San Juan Mountains has been examined by a number of studies using a combination of

remote sensing, isotopic analysis, back-trajectory tracking, and dust particle size analysis (Bennett and Depaolo 1987, Painter *et al* 2007, Neff *et al* 2008, Lawrence *et al* 2010). These studies provide direct and indirect evidences implying that dust deposited in the San Juan Mountains comes primarily from low deserts in northeastern Arizona and northwestern New Mexico.

Here, we use observations from the western United States to demonstrate that the variation of dust deposition on the snow is strongly linked to bare soil dynamics in dust-source regions. We focus on the dust-source area of the Little Colorado River basin in northeastern Arizona and the deposition of that dust in the San Juan Mountains of southwestern Colorado. Knowing the relationship between bare soil dynamics in the dust-source area and amount of dust in snow would improve our ability to predict magnitude and frequency of dust deposition worldwide, thus enabling more efficient management practices for water managers, land managers, and policy makers. In addition, this study is also important for land surface and earth system modelers interested in simulating climate effects from dust via vegetation dynamics.

2. Sites and methods

The dust-source area of this study is located in northeastern Arizona, the United States (figure 2). This area consists of mostly flat-lying region with elevation ranging from 1524 m in the low desert to over 2400 m in Black Mesa. A large portion of this area is covered by desert scrub, with widespread shrubs including four-wing saltbush (*A. canescens*), blackbrush (*C. ramosissima*) and shadescale (*A. conterfolia*). This area is characterized by mild winter and extremely hot summer, with an annual average precipitation of 160 mm, and a yearly mean temperature of 13 °C recorded in Tuba City, Arizona (figures 4(c) and (d) for precipitation). Precipitation comes primarily during two rainy seasons, with cold fronts coming from the Pacific Ocean during the winter and a monsoon associated with the moisture-bearing winds from the Gulf of Mexico in the summer (Adams and Comrie 1997). The monsoon season occurs towards the end of summer causing high precipitation during July–October. In the southwestern United States, strong convective winds, caused by excessive heating of the ground, along with synoptic events during the spring and early summer time cause frequent dust storms (Brazel and Nickling 1986).

Data on the amount of dust deposited in snow (including both wet and dry deposition) were obtained from dust-laden snow located in the San Juan Mountains, southwestern Colorado (figure 2). In this study area, two snow-study plots, one in the alpine zone at an elevation of 3719 m and one in the subalpine zone at an elevation of 3368 m, were set up to monitor snow and dust deposition beginning in winter of 2005 (Painter *et al* 2012a). During the same period, the change of vegetation cover in the dust-source area was characterized by the relative spectral mixture analysis (RSMA) technique of Okin (2007) using the moderate resolution imaging spectroradiometer (MODIS) surface reflectance data. For our

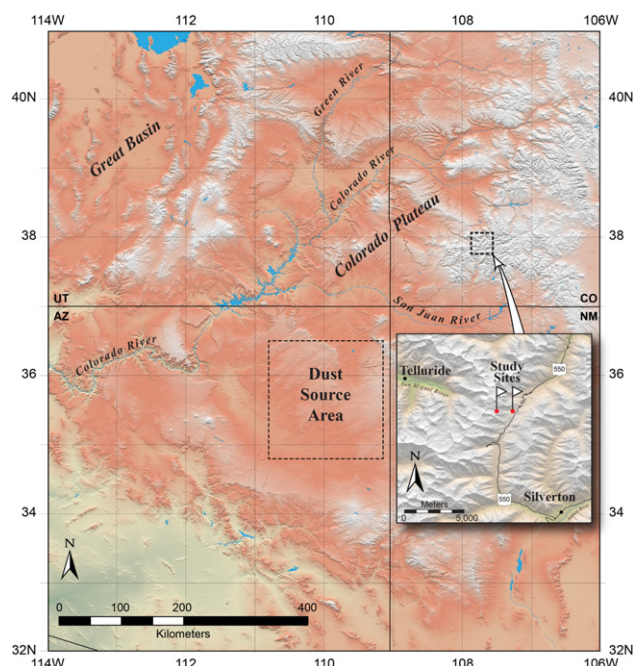


Figure 2. Location of the study area in the western United States. The area where dust-laden snow was collected is indicated in the overview inset. The dust-source area considered in this study is denoted by the dashed rectangular. In this map, mountain ranges are denoted by light, whitish colors and low deserts are represented by red colors.

purposes here, bare soil cover is considered the complement of total (green + non-photosynthetic) vegetation.

2.1. Dust on snow sampling

A study by Conway *et al* (1996) shows that dust particles are generally large enough that they are not entrained and washed to deeper layers during the process of snowmelt. Instead, they remain in their layer while overlying snow melts and percolates to below the dust. Therefore, the end-of-season dust deposition on the snow was determined by collecting dust samples (0.03 m × 0.05 m) just prior to snowpack depletion when dust from all events has converged at the surface. The actual date that was identified as prior to snowpack depletion varied from 13 May in 2009 (the earliest) to 20 June in 2011 (the latest), and the depth of snowpack for the end-of-season dust-deposition sampling ranged from 0.38 to 1.21 m. During the dust-sampling period, no additional dust-deposition events occurred after sampling the existing snowpack. A dust-deposition event was defined as any fresh deposition of mineral dust that is visible with the naked eye, either on the snowpack surface or in a snowpit wall as a layer within the snowpack (the visual approach can detect dust events containing as little as 0.1–0.2 g m⁻²). Dust and snow samples were weighed to 0.0001 g using an analytical balance in the laboratory. The final, end-of-season dust concentrations were reported as the dust concentration in milligram of dust per gram of snow, with an accuracy of 0.01 mg g⁻¹. More details on the experimental setup and observations may be found in Painter *et al* (2012b).

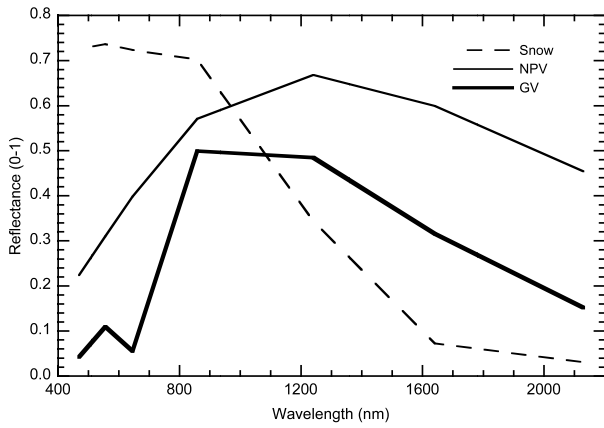


Figure 3. Reflectance spectra for green vegetation (GV), non-photosynthetic vegetation (NPV) and snow used as endmembers in relative spectral mixture analysis (RSMA). Redrawn from Okin (2007).

2.2. Remote sensing and statistical analyses

RSMA was applied to tiles h08v05 and h09v05 of MODIS surface reflectance data to obtain the dynamics of soil and vegetation cover in the dust-source area (Okin 2010). Unlike ordinary vegetation indices (i.e., the normalized difference vegetation index, NDVI) that provide information about green vegetation (GV), RSMA also quantifies changes in non-photosynthetic vegetation (NPV) cover, which can contribute substantially to the spectral signature of vegetation but is highly similar to spectra of the soil background and can therefore be difficult to quantify (Okin 2007, 2010). RSMA was applied to the 16-day MODIS 500-m nadir BRDF-adjusted reflectance (NBAR) product (MCD43A4) for the period of 2002–2011.

We initialize RSMA with a ‘baseline’ spectrum for each pixel in an image at a reference time, t_o . Here, t_o was set to be the MODIS-derived NBAR on 1 June (Day 153), 2002. In RSMA, the spectra of soil are assumed to vary spatially and to be strictly unknown. The spectra of GV, NPV, and snow are assumed to be constant across the scene and through time, with endmember spectra shown in figure 3.

RSMA indices, X , are calculated at each time step t_i by modeling the pixel reflectance ρ as a linear combination of the baseline spectrum, a GV spectrum, an NPV spectrum, and a snow spectrum:

$$\rho_{\text{pixel}}^{t_i} = (X_B^{t_i} + 1)\rho_B + X_{\text{GV}}^{t_i}\rho_{\text{GV}} + X_{\text{NPV}}^{t_i}\rho_{\text{NPV}} + X_{\text{Snow}}^{t_i}\rho_{\text{Snow}} + \varepsilon, \quad (1)$$

where $X_B^{t_i}$, $X_{\text{GV}}^{t_i}$, $X_{\text{NPV}}^{t_i}$, and $X_{\text{Snow}}^{t_i}$ are RSMA indices for baseline, GV, NPV, and snow, respectively, and ε is the residual error. $X_B^{t_i}$, $X_{\text{GV}}^{t_i}$, $X_{\text{NPV}}^{t_i}$, and $X_{\text{Snow}}^{t_i}$ must sum to zero. The RSMA indices can be either positive or negative and represent the changes in the fractional cover of that ground component relative to the reference time, t_o . Okin (2007) showed that the $X_B^{t_i}$ is equivalent to the ratio of the fractional cover of the soil at time t_i to the fractional cover of the soil

at time t_o . Here, the soil exposure index, $X_{\text{soil}}^{t_i}$, is calculated as one minus the sum of GV, NPV, and snow RSMA indices.

For each pixel, a time series of soil exposure anomaly, $X_{\text{soil,anom}}^{t_{\text{year,day}}}$, was calculated as:

$$X_{\text{soil,anom}}^{t_{\text{day,year}}} = X_{\text{soil}}^{t_{\text{day,year}}} - \frac{1}{n} \sum_{\text{year}=1}^n X_{\text{soil}}^{t_{\text{day,year}}} \quad (2)$$

where $X_{\text{soil}}^{t_{\text{day,year}}}$ is the soil exposure index for a *day* in a *year*, and n is the number of years in the analysis (here, the analysis covers 2005–2011, $n = 7$). In practice, *day* was the first day in the 16-day composite period for the MODIS NBAR data. Here $X_{\text{soil,anom}}^{t_{\text{year,day}}}$ was calculated for $\text{day} = 1$ (1 January) to 153 (1 June), the period during which dust deposition to snow cover generally occurs in the Colorado River Basin (Painter et al 2012b).

The Spearman rank correlation between $X_{\text{soil,anom}}^{t_{\text{year,day}}}$ (for each $\text{day} = 1-153$) and end-of-season dust concentration in snow was calculated and tested for significance at the $\alpha = 0.05$ level (critical value = 0.714 for $n = 7$). The use of non-parametric rank correlation was justified here because (1) the small n indicates that assumptions of normality cannot be met, and (2) no assumption of linearity between RSMA-derived bare soil dynamics and dust deposition in snow can reasonably be made (Sokal and Rohlf 1995).

3. Results

During the period of 2005–2011, both the magnitude of dust loading on snow and the number of dust-deposition events varied dramatically (figures 4(a) and (b)). The end-of-season dust concentration varied from 0.26 mg g⁻¹ in 2005 to 4.22 mg g⁻¹ in 2009, corresponding to the number of dust-deposition events of 4 and 12, respectively. The extreme dust deposition observed in 2009 also corresponds to a driest year on record in northeastern Arizona and southeastern Utah (figures 4(c) and (d)). Despite the mean precipitation being rather large for 2010 in the dust-emitting area, a second largest dust load of 3.51 mg g⁻¹ was observed in the dust-deposition area, which is presumably related to the substantially dry period from April to June in the dust-source area (figure 4(d)). The fact that the end-of-year dust concentration in the snow does not perfectly coincide with the number of dust events in that year indicates that these two factors are not directly related. In particular, different dust-deposition events can have more or less mass associated with them. The possible reasons for this are many: differences in emission rates/duration in the source area, differences in sorting/scavenging during transport, differences in transport pathway, differences in scavenging rate/duration for wet deposition events, differences in the exact spatiotemporal distribution of snowfall, etc.

In the dust-emitting area, remote sensing retrievals of soil and vegetation cover dynamics show that strong positive anomalies of bare soil exposure were found in years 2009 and 2010, whereas negative soil anomalies were observed in years 2005–2007, corresponding to a magnitude of dust

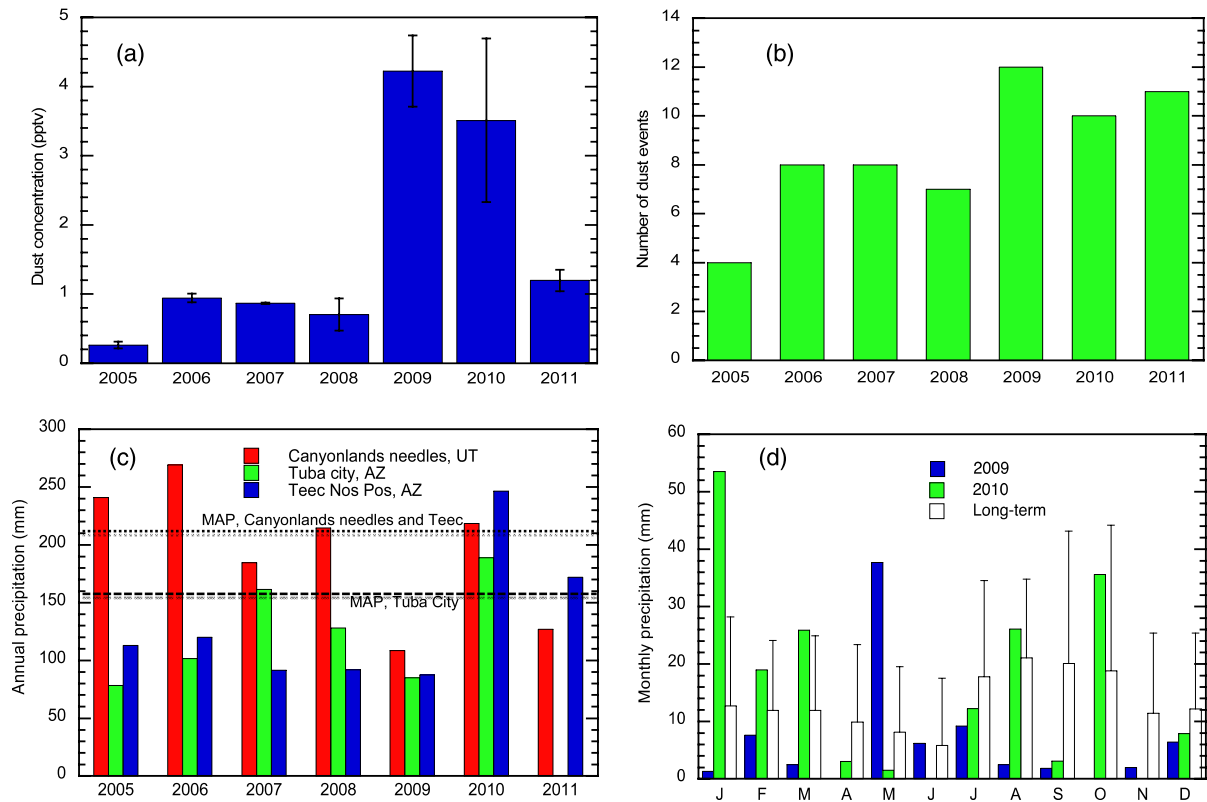


Figure 4. (a) End-of-season dust concentration in snow measured at two study plots located in the snow mountains of southwestern Colorado, error bars are one standard deviation of dust concentration obtained from the alpine and subalpine study plots ($n = 2$). (b) Number of dust-deposition events to snow cover in dust-deposition area, southwestern Colorado. Note that the number of dust-deposition events for a year counted from previous winter to the end of snow season. (c) Mean annual precipitation (mm) for weather stations located in the dust-source area and long-term mean annual precipitation (MAP, dashed lines), during the period of 2005–2011. (d) Monthly precipitation for two extreme dust-deposition years (2009 and 2010) in the dust-source area, Tuba City, Arizona. Precipitation data were derived from the Western Regional Climate Center (www.wrcc.dri.edu/) with >45 years for long-term averages. Mean annual precipitation for Tuba City in 2011 was not available. Error bars are one standard deviation.

deposition of high to low in the dust-deposition area, respectively (figures 5(a) and 4(a)). Looking in more detail, 2009 had the highest dust deposition-on-snow with annual precipitation in 2009 also among the lowest. Nonetheless, the May 2009 rainfall (figure 4(d)) is relatively high. To investigate this contradiction, we investigated the dates of the 12 dust-deposition events that contributed to the loading in 2009 (data available at: www.snowstudies.org/CODOS/dustlog.html) occurred prior to May (the last was 25 April, 2009). Thus, we conclude that the relatively high precipitation in May 2009 had no effects on the dust production and observed high dust loading in the snow.

The analysis of $X_{soil}^{day,year}$ revealed areas in which clusters of contiguous pixels were positively and significantly correlated with end-of-season dust loading in the dust-deposition area (figure 5(b) for Julian day of 121 as an example). A majority of these clusters are found in the northwestern and southeastern part of the dust-source area, including the relatively low-elevation portion of the Black Mesa. In the dust-source region, the number of significantly correlated pixels was low during the winter but increased gradually from early to late spring to account for 15% of the MODIS pixels in a box drawn around the potential source area

on day 121 (1 May, figure 6). The time when there were the most significantly (positively) correlated pixels corresponds well with the seasonality of dust-deposition events in the San Juan Mountains.

4. Discussion and conclusions

Evidence from dust-source areas (i.e., areas known to produce dust under some circumstances) indicates that many factors could impact the interannual variability of dust emission from the source area and the deposition of this dust in the mountain snowpack. Conditions of the soil surface that impact this variability include the threshold for particle transport (i.e., soil erodibility associated with soil moisture content) and soil erodible fraction (i.e., soil cover and fraction of soil that is uncrusted) (e.g., Field et al 2010, Pierre et al 2012). Meteorological conditions that impact this variability include the intensity of the wind (i.e., its erosivity), the dust transport pathway, and deposition conditions (Prospero and Lamb 2003). In the dust-emission area of this study, the emission of dust may be also significantly affected by the origin of winds (e.g., convective versus synoptic) (Brazel and Nickling 1986).

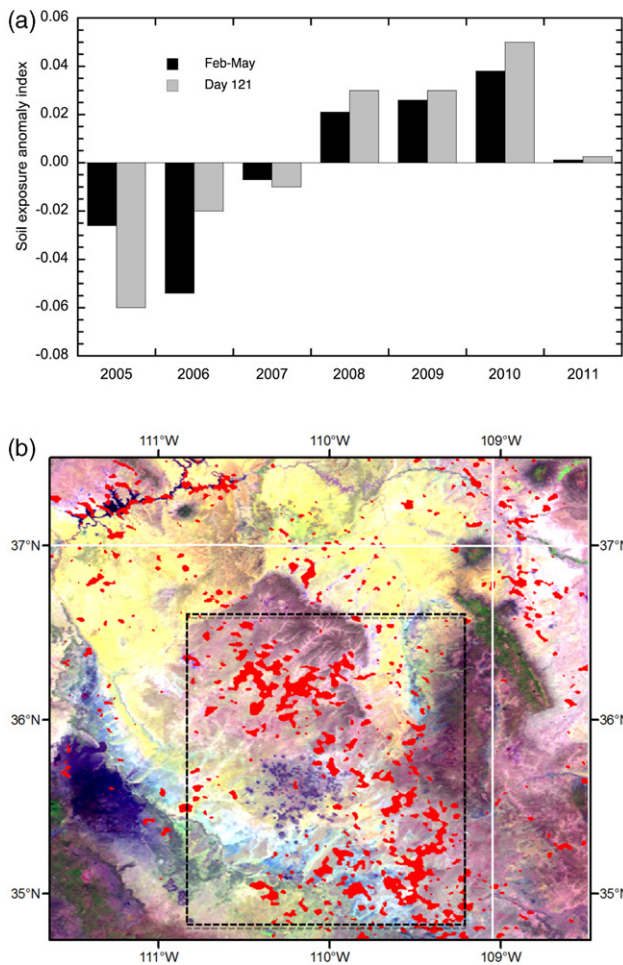


Figure 5. (a) RSMA-derived soil exposure anomaly for the dust-source area in northeastern Arizona (encompassed by the dashed rectangular in figure 2), averaged for the period of February–May and day 121 for 2005–2011. (b) Spatial distribution of pixels with significant positive correlations ($\alpha < 0.05$) between soil exposure anomaly and end-of-season dust concentration in snow mountains, southwestern Colorado. The results shown here are for correlations on day 121, the date with highest number of significantly correlated pixels. This image covers part of MODIS tiles h08v05 and h09v05 and was created by overlaying the RSMA baseline image with significant pixels in red. In the baseline image, red is MODIS band 6 (shortwave infrared), green is MODIS band 2 (near infrared), and blue is MODIS band 3 (blue).

Because snow is neither common nor long lasting in the dust-source area (e.g., the low deserts) during the spring, soil exposure mirrors the cover of total vegetation, determined by the total contributions of GV and NPV. In the northeastern Arizona, Okin (2007, 2010) revealed that the dynamics of soil exposure may be attributed to the intra-annual variations of NPV, rather than the GV, which varies only slightly through out the year. Here, we investigate how the variation of a remotely sensed surface property (bare soil) correlates with the variation of dust deposition on the snow in the Western United States. It should be noted that the dust-source area defined in this study is broad and also includes Black Mesa, a high-elevation, mostly incised landscape that contains much rock-covered or permanently (but sparsely) vegetated

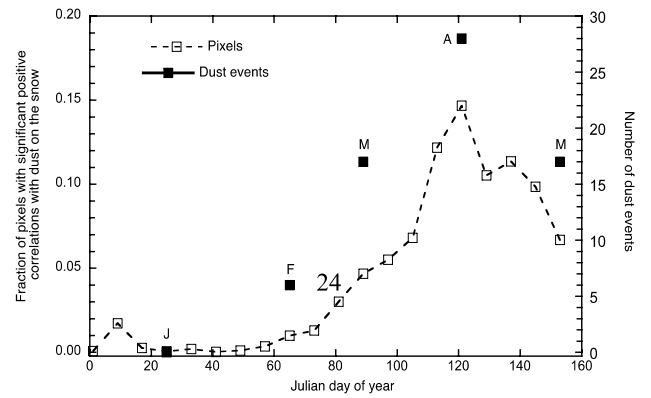


Figure 6. Fraction of pixels (day 1–153) exhibiting significant positive correlations ($\alpha < 0.05$) with end-of-season dust concentration in the snow for the dust-source area (area shown in dashed rectangular of figure 1) for years 2005 through 2011. Squares indicate the first date of 16-day composites MODIS data (produced every eight days). Day 121 has the highest proportion of pixels with positive correlations. This figure is overlaid with the monthly average number of dust-deposition events (denoted as black squares) in the San Juan Mountains from January to May, represented as the last day of each month in the x-axes as Julian day of year. Dust-deposition events data were obtained from the Center for Snow and Avalanche studies (www.snowstudies.org), Silverton, Colorado.

terrain that appears to produce insignificant dust compared with the low deserts (Reynolds 2013). Therefore, we do not argue that all pixels that exhibit statistically significant correlations with end-of-season dust concentration produce dust that is deposited in the dust-deposition area. Nonetheless, the existence in known dust-source areas of statistically significant correlations between a surface property (soil cover) and dust deposition suggests that surface conditions are a primary control on the variability of dust deposited in the snow in the mountains of southwestern Colorado.

There are other environmental factors that likely also contribute to the variability in dust deposition in the snowpack, including variations in soil threshold shear velocity due to wetting (e.g., Pierre *et al* 2012) or soil disturbance (e.g., Belnap 1995) and meteorological conditions (e.g., Brazel and Nickling 1986). To the extent that these are not related to bare soil cover as it is sensed through RSMA analysis of MODIS data, though, these factors must be in the portion of the variance that is not explainable by bare soil cover from RSMA. The extensive significant correlations between RSMA-derived bare soil cover and dust deposited in the mountain snowpack is a strong argument for the importance of bare soil cover as a primary control on dust emission from the region. The results need not have turned out this way; if meteorological conditions, soil moisture, or soil disturbance were the dominant controls, the signal from the bare soil cover would not be as strong as it is because RSMA is insensitive to these parameters.

Our assertion that (remotely observable) bare soil cover is a primary control on dust emission and deposition in distal snowpack is supported by a number of features of our findings. First, we find that the per cent of pixels in the dust-source area (up to 15%, figure 6) that exhibit correlations

between deposited dust and soil cover anomaly is much greater than that predicted by chance (5% at $\alpha = 0.05$ level). Moreover, we find many fewer pixels than would be predicted by chance exhibiting correlations with snow dust loadings in the first quarter of the year, when there are few dust-deposition events, supporting the strength of the correlations during the dust emission/deposition season. Second, we find that spatially contiguous clusters of pixels in known source areas consistently exhibit this correlation because source areas are larger than the 500-m pixels used here (e.g., figure 5(b)). Spurious correlations due to chance would not be expected to exhibit spatial contiguity. Third, we find that the peak correlation occurs when it would be expected to, that is, during the time of year when much dust is produced (figure 6). Regionally derived dust does not likely spend more than a few days in transport and thus at the temporal scale examined here, there should be no lag between dust deposition and bare soil dynamics. Fourth, we find that the observed extreme dust deposition in 2009 and 2010 (figures 4(a) and (b)) also had the highest integrated springtime soil cover anomaly (figure 5(a)). This observation suggests that soil cover, which should be controlled, at least in part, by precipitation and consequent vegetation growth, is the dominant surface control on dust emission. Fifth, we find that precipitation records from stations near the dust-source area indicate poor conditions for vegetation growth in the first part of the year for 2009 and 2010 when the greatest dust deposition occurred. Poor rainfall likely leads to little vegetation growth resulting in increased bare soil cover and improved conditions for dust emission.

This argument presupposes that geographic and meteorological conditions already exist that allow deposition of dust in the mountains from a certain source region, but this supposition is already known to be true. Dust emission from certain specific parts of our study area and deposition to the mountains of the Upper Colorado have been shown through remote sensing of large dust plumes and back-trajectory analysis of known dust-deposition events on snow (Painter *et al* 2007). Furthermore, Steenburgh *et al* (2012) show that the meteorological conditions that transport dust to the nearby Wasatch Mountains in north and central Utah from 2001 to 2010 remain relatively consistent from year to year. Although not directly applicable to dust deposited in the mountain snow cover of southwestern Colorado, the meteorological conditions that produce dust in that study are likely similar to those that produce dust in our study area because they are due to the same large-scale synoptic events (i.e., the Nevada low).

Soil cover likely covaries with other surface properties that could impact dust emission. For instance, grazing could reduce vegetation cover, thereby increasing soil exposure, while also disturbing the soil. Soil disturbance leads to both lowered threshold for the initiation of saltation (Belnap *et al* 2007) and reduced area of soil crust (Baddock *et al* 2011), both of which would increase dust emission. Our analysis cannot rule out some processes that covary with soil cover that cause increased dust emission in the dust-source area, but the precipitation record does suggest that years that are very dry in the first quarter of the year (e.g., 2009 and 2010) are also years

with high dust deposition and years with relatively wet first quarters (e.g., 2005 and 2007) have negative soil anomalies and relatively little dust deposition.

Projected climate changes will likely bring more frequent and sustained drought to the southwestern United States (Seager *et al* 2007). The projected increase in aridity in the southwestern United States is expected to cause reductions in vegetation cover, therefore leaving greater bare surface for wind erosion, and accelerated rates of dust emission (Seager *et al* 2007, Munson *et al* 2011). The projected increase in temperature in this region is likely to decrease average soil moisture, and strongly amplify dust emissions through its direct impacts on threshold shear velocity and indirect influences on vegetation cover, fire regimes, and hydrophobicity of soils (Ravi *et al* 2007, Ryan *et al* 2008, Pierre *et al* 2012). Dust emission may be further amplified by human activities such as agriculture, grazing, and resource exploration in arid and semiarid regions (Neff *et al* 2008, Field *et al* 2010). Our work suggests that in our study area, change of bare soil cover associated with variations in precipitation may lead to an order of magnitude greater of dust deposition (e.g., year 2009 versus 2008) to the mountains. The extreme dust deposition in 2009 thus may represent a vision into the future pattern of dust forcing snowmelt dynamics in the Colorado River Basin, subject to the influences of climate change on snowfall and rainfall.

While our study shows that the surface conditions (e.g., bare soil exposure) in the dust-source area have a primary impact on the variability of dust loading in the mountain snow cover of southeastern Colorado, we cannot rule out the potential impacts of other factors, such as meteorological conditions, on the emission of dust in the source area. In this study, dust samples were collected once during the dust-deposition season whereas the San Juan Mountains receive multiple dust-deposition events annually, future studies that look at finer-scale temporal linkages between bare soil and dust loading may be needed, supported by additional dust profiles throughout the years.

Acknowledgments

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