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Dust effects on snowpack melt and related ecosystem processes are secondary to those of forest canopy structure and interannual snowpack variability

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ABSTRACT

Dust deposition lowers the albedo of snow and can significantly alter snowpack energy balance. Investigation of aeolian dust deposition in the mountains of the western U.S. has shown that these effects advance the timing of snowpack melt and spring runoff across much of the region. These studies have primarily focused on alpine snowpacks with little to no overstory vegetation. To evaluate the impacts of aeolian dust on ecohydrological processes in forests, we conducted a manipulative experiment in a subalpine conifer forest in Utah's Wasatch Mountains. During the spring of 2010–2012, we added dust to the snow surface in forested plots every 1 to 2 weeks, roughly doubling the natural dust loading. We then measured snowpack ablation in control and dust addition plots, along with below-snowpack and warm season soil temperature (T_{soil}), soil water content (θ), litter decomposition rate (D), soil respiration rate (R_s), and tree xylem water potential (ψ). Differences in ablation between control and dust addition plots were similar in magnitude to differences associated with the canopy structure of the forest. Seasonal patterns in T_{soil} and θ were similar between dust treatments and canopy structure groups. D, R_s, and ψ varied little between dust treatments, but there were significant differences between years. During our 3-year study, an unusual level of interannual variability in snowfall had the greatest effect on the soil environment and ecosystem processes. The effects of aeolian dust on snowpack mass and energy balance in our forest were slightly smaller than those associated with canopy structure.

KEY WORDS ecohydrology; soil respiration; decomposition; snow hydrology; winter biogeochemistry

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INTRODUCTION

Dust and other impurities lower the albedo of snow and have additional indirect effects on the energy balance of snow-covered and ice-covered land surfaces (Warren and Wiscombe, 1980; Hansen and Nazarenko, 2004). During the spring, solar energy absorbed by particles near the snow surface can hasten the warming and melting of the snowpack (Conway *et al*., 1996; Painter *et al*., 2007; Gleason *et al*., 2013). Recent studies have demonstrated that deposition of aeolian dust on mountain snowpacks leads to a significantly earlier timing of snowpack melt and seasonal water runoff in the hydrologic basins of the western U.S. (Painter *et al*., 2007; Skiles *et al*., 2012). Studies that model the effects of dust on snowpack dynamics have sometimes included forested areas, but experiments directly examining the effects of dust deposition on ecological processes have, to date, been limited to alpine areas where there is no vegetation canopy above the snowpack (Steltzer *et al*., 2009).

Snowpack energy balance in forested areas differs from that in open, alpine areas. A fraction of incoming shortwave (solar) radiation is intercepted by and warms the canopy, which then increases the emission of longwave (terrestrial) radiation towards the snow surface. Snow is an efficient absorber of longwave radiation, and this radiation becomes an important energy source for ablation in below-canopy environments (Link and Marks, 1999a, 1999b; Koivusalo and Kokkonen, 2002; Link *et al*., 2004; Pomeroy *et al*., 2009; Varhola *et al*., 2010). Dust deposition lowers the shortwave albedo of snowpacks regardless of the presence of a canopy, but it does not appreciably enhance longwave absorption by snow (Warren and Wiscombe, 1980; Painter *et al*., 2007). The efficacy of dust in perturbing snowpack energy balance below a canopy should therefore depend on the relative contributions of shortwave and longwave radiation, which are strongly influenced by canopy structure and radiative transfer (Link and Marks, 1999a; Sicart *et al*., 2004; Ellis *et al*., 2011; Lawler and Link, 2011). At present we know

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of no manipulative studies that have addressed the effects of dust deposition on snowpack melt in forested areas.

Reduced snowpack and earlier melt timing are associated with a variety of effects on ecosystems. Active microbial communities are present beneath seasonal mountain snowpacks, and their activity is tied to below-snowpack temperature and water availability. The melting of spring snow triggers the turnover of these communities and an associated flush of nutrients (Brooks *et al*., 1998; Jaeger *et al*., 1999; Lipson *et al*., 1999). The spring snow melt also marks the beginning of a more physiologically active period for many organisms, and changes in the timing of melt can alter the timing of emergence, greening, and flowering in alpine plant communities (Steltzer *et al*., 2009) and activity of birds and animals (Inouye *et al*., 2000; Ozgul *et al*., 2010). Warm season activity by plant and soil communities in snow-dominated ecosystems depends heavily on snowmelt water (Brown-Mitic *et al*., 2007; Litaor *et al*., 2008; Riveros-Iregui and McGlynn, 2009), and differences in snowpack size and melt timing can have a significant effect on forest productivity (Molotch *et al*., 2009; Tague *et al*., 2009; Hu *et al*., 2010). Perturbations to snowpack ablation by dust events may therefore have a significant effect on a variety of ecosystem processes.

Dust deposition has changed since the settlement of the western United States, largely because of human-driven land use and land cover change (Neff *et al*., 2008; Painter *et al*., 2010; Ballantyne *et al*., 2011). Recent studies have revealed declining trends in snowcover extent, duration, and snowpack size in the region over this time period (Hamlet *et al*., 2005; Mote *et al*., 2005; Mote, 2006; Dyer and Mote, 2007). These trends in the timing and magnitude of snowpack ablation are thought to be responsible for shifts towards earlier spring runoff timing in the hydrologic basins of the western U.S. (Dettinger and Cayan, 1995; McCabe and Clark, 2005; Regonda *et al*., 2005; Stewart *et al*., 2005; Hamlet *et al*., 2007). The snowpack and streamflow changes reported in the interior western U.S. (Clow, 2010; Nayak *et al*., 2010; Harpold *et al*., 2012) are consistent with regional and global trends in earth surface temperature change but may also be attributable, in part, to the effects of aeolian dust deposition on mountain snowpacks (Painter *et al*., 2010). According to model projections, increasing trends in aridity and temperature in the western U.S. will continue and intensify in the coming century (Brown and Mote, 2009; Seager and Vecchi, 2010; Kapnick and Hall, 2012). These trends bring a high likelihood of widespread vegetation change and greater aeolian dust fluxes (Westerling *et al*., 2006; Logan *et al*., 2010; Anderegg *et al*., 2011; Munson *et al*., 2011), which may act as a positive feedback for further hydroclimatic changes in the region.

Although numerous studies suggest that increased dust deposition in the western U.S. will lead to hydrologic and ecological change, few direct experiments have been performed. Given that changes in dust deposition are concomitant with changes in temperature, aridity, vegetation cover, and other factors, it is critical that the mechanisms of ecosystem responses to dust deposition be investigated. We added dust to the snowpack beneath a subalpine conifer forest in Utah and measured resulting changes in snow water equivalent (SWE), snow ablation, and the soil environment, including soil temperature (T_{soil}) and soil water content (θ). We then monitored the response of vegetation and soil biological processes, including xylem water potential (ψ), soil respiration flux (R_s), and litter decomposition rate (D), to this snowpack manipulation. We hypothesized that dust addition would increase the rate of spring snowpack melt, leading to earlier snow melt, decreases in warm season θ, and changes in the seasonal pattern of T_{soil}. We expected responses from vegetation and soil biological processes that would follow the timing and magnitude of changes θ and T_{soil}. This experimental design and fortuitous timing allowed us to assess the role of within-forest differences in canopy structure and high interannual variability in snowpack dynamics.

METHODS

Site description

In the spring of 2010, 2011, and 2012 we conducted a snowpack manipulation at a Rocky Mountain subalpine forest to measure the impact of dust deposition on snow ablation below a conifer canopy. The study took place in a mature conifer forest on a south facing slope (slope = 21°, aspect = 202°) at 2895 m (40°, 36′N, 111°, 35′W) in the Wasatch Mountains near Salt Lake City, Utah. Dominant conifer species in this forest were *Abies lasiocarpa* (subalpine fir) and *Picea engelmanii* (Engelmann spruce), and there were small patches of the deciduous *Populus tremuloides* (quaking aspen). This forest was intentionally chosen for its patchy, open canopy structure and southern aspect, which we assumed would allow significant transmission of shortwave radiation through the canopy.

Beneath this canopy, we delineated 10 × 60 m study plots with long edges oriented parallel to the direction of the site slope. In 2010 we established a pilot snowpack manipulation consisting of one control and one dust addition plot. At this stage we attempted to control for canopy structure by measuring stem density of the study forest and locating our study plots in areas of the forest with similar density. However, as we added replicates, we decided to additionally account for canopy structure using hemispherical photography (described below). In 2011 we added two replicates to each treatment for a total of three

10 × 60 m plots per treatment. Control and dust addition treatments were randomly assigned to the plots. In October 2009, we installed a weather station in a clearing outside the study forest. We also installed six soil moisture sensor profiles (CS-616, Campbell Scientific, Inc., Logan, UT, USA; EC-5, Decagon Devices, Inc., Pullman, WA, USA) and two soil temperature sensor profiles (Decagon EC-5) in each treatment. In September 2010 we added an additional sensor profile for θ (Campbell CS-616) and for T_{soil} (Campbell CS-107), for a total of seven θ and three T_{soil} profiles per treatment. The sensors in each profile were installed at 5, 20, and 60 cm below the top of the mineral soil horizon. Thirty-second readings of T_{soil} and θ were logged and then averaged every half hour with Campbell Scientific 23X dataloggers. See Figure 1 for a schematic of the experimental design.

Snowpack dust addition

Dust provenance. For the 2010 pilot project, we collected dust from the Chinle–Moenkopi formation of the Colorado Plateau. This geologic formation is a source for aeolian dust for some areas of the southern Rockies (Neff *et al*., 2008; Lawrence *et al*., 2010). After the pilot project was complete, however, our colleagues established that the Wasatch Mountains receive significant amounts of dust from Great Basin regions to the south and west (Steenburgh *et al*., 2012). On the basis of this new understanding of Wasatch dust sources, we changed the dust source for the remainder of the study. The Milford Flat fire near Filmore, UT, in the summer of 2007 was the largest wildfire in Utah history, and the burned area has become a recognized source of the windblown dust deposited in the Wasatch Mountains (Hahnenberger and Nicoll, 2012; Miller *et al*., 2012; Steenburgh *et al*., 2012). In March of 2011 we collected dust from drifts of wind-deposited material along a roadway through the Milford Flat fire scar. Although this material had different spectral characteristics than the Colorado Plateau dust, it visibly darkened the surface of the snowpack when applied. The material collected from both dust sources was sifted to < 500 μm. This size threshold is larger than the typical size class for aeolian dust (Lawrence *et al*., 2010) but produced material that could be easily scattered across our 10-m-wide plots.

Dust application. During the spring, dust was scattered by hand from the edge of the dust addition plots on to the surface of the snowpack. Care was taken to avoid trampling the snowpack inside the plots. We timed these dust additions to follow new snow events and, when possible, to precede clear, sunny weather. A new dust addition occurred every 1 to 2 weeks at times that maximized the exposure of the dust on the snowpack surface to shortwave radiation and thus its effect on the snow melt rate. We anticipated that six artificial dust events per year, at a loading rate at roughly 5 g/m^2 would more than double the annual ambient dust loading observed in our region (Lawrence and Neff, 2009). We applied dust six times in 2010 and 2011 and only four times in 2012 due to a smaller snowpack and early spring snowmelt in that year. To verify that dust addition had increased the amount of particulate matter in our snowpack above ambient levels, we collected cores of the full snowpack column from all plots once the final dust application was made each year. In 2011 and

Figure 1. Schematic of the study forest, including the location of all snowpack manipulation plots. The weather station was located in the clearing to the northwest of the plots.

2012 we also measured ambient particulate matter loading in a clearing near our forest. In 2011 the clearing measurement was made by excavating a full snowpit on 23 May and sampling the entire snowpack in 10-cm increments. For the rest of the spring after this full snowpit collection, we collected surface cores ($n = 3$) on a storm board following each natural dust event, and the dust mass in these cores was added to the total dust loading from the snow pit. In 2012, three full snowpack cores were collected in the clearing on the same day as those collected in the canopy, and there were no further dust events after this collection. Snow cores and pit samples were thawed and filtered through weighed glass fibre filters (Whatman Grade GF/A, GE Healthcare Bio-Sciences, Pittsburgh, PA, USA), and the filters were oven dried and weighed to determine total particulate matter loading at each location. From these filters, we removed particulate matter that was clearly forest litter (needles, bark, scales, and such) and weighed it separately.

Snow measurements. At six locations in each plot ($n = 18$ in each treatment), SWE of the snowpack was measured prior to each dust addition and on a roughly weekly schedule once ablation began. Measurement locations were marked and remained the same (±3 m) for the duration of the experiment. SWE measurements were made using a Federal aluminium tube snow sampler (Union Forge, Yakima, WA, USA). Precipitation and SWE data from the Brighton SNOTEL site (Site 366, USDA, Natural Resources Conservation Service, http://www.wcc.nrcs.usda.gov/snow/) were used for some of our analyses. This station was located at the edge of a clearing < 2 km from our study forest, at an elevation of 2670 m on a similar aspect to our study site.

Ecosystem process measurements

Below-snow soil respiration. During spring of 2011 and 2012 we measured R_s below the snowpack using the diffusion gradient method outlined in Sommerfeld *et al*. (1996). Nine 10-cm-diameter stainless steel mesh gas inlets were placed on the soil surface before the snowpack developed in control and dust addition treatments (18 inlets total). These inlets were routed to a central gas collection location between the plots using 0.64-cm-diameter tubing (Type 1300, Synflex Specialty Products, Mantua, Ohio). Collection tubes were capped with stainless steel gas-tight removable fittings (Swagelok Co., Solon, Ohio, USA). At sampling time, tubes were uncapped and attached to a small gas pump (NMP850, KNF Neuberger, Inc., Trenton, NJ, USA) via an inline flowmeter (Gilmont Instruments, Barrington, IL, USA). A volume of gas equal to the volume of the tubing was pumped away, and the pump was then isolated from the tubing. The gas in the tube was then sampled using a syringe (Becton, Dickinson and Company, Franklin Lakes, NJ, USA) through a septum (Hamilton Co., Reno, NV, USA) upstream of the pump and transferred to a pre-evacuated glass vial (Labco Exetainer, Labco Ltd., Lampeter, Ceredigion, UK). Three samples of air were collected using the syringe above the snowpack on each sampling date. Upon return to the laboratory, the CO_2 mol fraction in these samples was measured by injecting 0.5 ml of gas into a closed-loop infrared gas analyser system (LI-7000, Li-Cor Biosciences Inc., Lincoln, NE, USA; see Moyes *et al*., 2010). Soil respiration rate was calculated using Fick's law with adjustments for snowpack properties by

$$F = \rho_a \eta \tau D \frac{\mathrm{d}C}{\mathrm{d}z}$$

where ρ_a is the molar density of air (adjusted for temperature and pressure), η and τ are the porosity and tortuosity of the air-filled snowpack, respectively, D is the molecular diffusivity of CO_2 in air (adjusted for temperature and pressure following Massman, 1998), and C is the mole fraction of CO_2 at height z (see Bowling and Massman, 2011).

Warm season respiration. During the snow-free season we measured R_s from polyvinyl chloride collars roughly twice per month using a Li-Cor 6400 infrared gas analyser with a 6400-09 soil chamber attachment. In 2010, the unreplicated pilot plots were measured ($n = 10$ locations per treatment), and in 2011 and 2012 four measurements were made in each of all six plots ($n = 12$ locations per treatment). Collars were inserted about 2.5 cm into the soil surface in an evenly spaced line down the middle of each plot and were moved by 1 m in a random direction at the start of each new season. Measurements of T_{soil} at 5 and 15 cm depth (thermocouple probe, Omega Engineering Inc., Stamford, CT, USA) and surface θ were taken at each respiration collar at the same time (Campbell Scientific CS-620 probe).

Warm season xylem water potential. In spring 2010 we selected 18 mature subalpine fir trees in the pilot plots ($n = 9$ per treatment) and measured predawn and midday ψ using a pressure chamber (PMS Instrument Co., Albany, OR, USA) roughly twice per month until the fall. In 2011 we added three subalpine fir saplings (diameter at breast height (DBH) < 2 cm) in each plot for measurement of ψ ($n = 9$ per treatment). In 2011 and 2012 we measured predawn and midday ψ in these saplings on the same schedule as R_s measurements. We continued in these years to measure a subset of the mature subalpine firs ($n = 5$ per treatment) but less frequently than in 2010. We did not control for the horizontal area of the rooting zone of these trees, and the roots of measurement trees may have extended beyond our plot boundaries (Day *et al*., 1989).

Litter bag mass loss (decomposition rate). In fall of 2010, we collected needle litter from canopy conifers at the site on tarps and oven dried. Five grams of litter was then sewn into nylon and fibreglass mesh litter bags (0.2 mm nylon mesh bottom, 1.7 mm fibreglass screen top). On 15 October

2010, at 36 locations in the study forest (18 per treatment), we placed a group of five litter bags on the forest floor and secured them with metal staples ($n = 90$ bags per treatment). From the time of placement until spring 2013, we returned to each litter bag group immediately following spring snowmelt and in late fall (~15 October) each year to collect one bag per location. Collected bags were placed in a drying oven for 48 h, and decomposed litter was carefully removed from the bag and weighed. Bags that were disturbed or damaged by animals ($n = 26$) were excluded from analysis.

Mass loss was described using an exponential decay model with two pools, one fast and one slow cycling (Adair *et al.*, 2008; Harmon *et al.*, 2009). This model took the form

$$L_t = L_{0f}e^{-\lambda_f t} + L_{0s}e^{-\lambda_s t}$$

where L_t is the fractional litter remaining at time t, L_{0s} and λ_s are the initial fraction and decay constant of the slow-cycling litter pool, λ_f is the decay constant of the fast-cycling litter pool, and L_{0f} is the initial fraction of the fast-cycling pool and is defined as $1 - L_{0s}$. We fit this model to our data using the nonlinear least squares method (Adair *et al.*, 2010).

Hemispherical photos

On several dates in 2012, we took hemispherical photographs of the canopy at all SWE measurement, litter bag, and warm season R_s locations, and at each soil sensor profile. For each photo, the camera tripod was adjusted to 1 m above the snow or soil surface, the camera lens was levelled, and upwards looking photos were taken with a circular fisheye lens (8 mm F3.5 EX DG Circular Fisheye, Sigma Corporation, Kanagawa, Japan). To capture the with-leaves and without-leaves canopy structure, we took photos at SWE measurement locations on 24 April (no leaves), at soil profiles and litter bag/respiration locations on 17 July (after aspen leaf out) and again at soil profiles on 17 October (after aspen leaf fall). We analysed each digital photo using Gap Light Analyzer v2.0 software (Frazer *et al.*, 1999). For each photo, this software calculates a value of canopy openness, the percent of a 180° sky view not occupied by canopy, and direct-beam transmissivity, the percentage of above-canopy radiation transmitted to the forest floor. The size of our ψ measurement trees and their variable rooting area prevented meaningful characterization of canopy structure above them.

Statistical analysis

We compared the effect size of dust versus canopy structure on snowpack ablation by fitting a simple statistical model to our data. We modelled the change in SWE between one measurement date and the next as a function of incoming solar radiation (measured as PAR), air temperature, and new snowfall (using SWE measured at the Brighton SNOTEL site). The basic form of this model was

$$dSWE_{it} = \beta_0 + \beta_1 AirT_{it} + \beta_2 Snow_{it} + \beta_3 Pin_{it} + \varepsilon_{it}$$

where $dSWE_{it}$ was the change in SWE measured at location i and time t, $AirT_{it}$, Pin_{it}, and $Snow_{it}$ are the integrated air temperature, incoming solar radiation, and snowfall measured at time t, respectively, $\beta_{0...3}$ were the intercept and regression coefficients for these independent variables, and ε_{it} was the residual error. We fit this model to our SWE measurements using least squares regression. Note that some of these independent variables were correlated, and thus, fitted regression coefficients do not accurately describe their relative importance for snowpack ablation. We expected the influence of these independent variables to vary according to treatment and canopy structure, so we also fit the data to variations of the model that included interactions for treatment (control or dust addition), canopy radiation transmission (high or low), and canopy openness (open or closed).

We used a multilevel linear model with sample date as a random variable (a repeated measures design) to test for differences in SWE during the accumulation period, which we defined as the first four SWE measurement dates of the field season. We compared differences in SWE between treatments and canopy groups with this technique. Similar multilevel model tests were used for comparisons of R_s, T_{soil}, θ, and xylem ψ. A similar multilevel model was used to test for interannual differences in these variables with measurement location as the random variable because the same locations were measured in all years. This was followed by post-hoc analysis (Tukey's honest significant difference (HSD)), and we report post-hoc test results ($\alpha = 0.05$) to indicate significant differences between the means of each year.

RESULTS

The Wasatch Mountains experienced three very different winters during the years of our snowpack manipulation experiment (Figure 2). In 2011, this region had a near

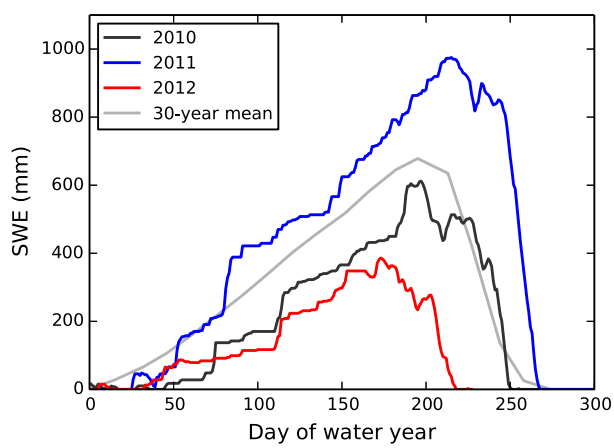

Figure 2. Snow water equivalent at the Brighton SNOTEL site, located about 2 km from our study forest, for the study years 2010 to 2012. The long-term mean for the site was calculated at 2-week intervals for the years 1971–2000 and is plotted for reference in grey.

record-breaking large snowpack, and in 2012 it had a near record-breaking small snowpack. The 2010 snowpack was intermediate with peak SWE similar to the long-term average. These differences allowed us to compare our snowpack and ecosystem process measurements between widely contrasting years.

Snowmelt manipulation

Dust addition visibly darkened the surface of the snowpack and successfully increased the load of particulate matter in the snowpack beyond the ambient snowpack dust load at the site. We measured the total particulate content of the snowpack in an adjacent clearing (no canopy present) and in control and dust addition plots and found that our dust additions roughly tripled the mass of particulates found in the clearing and doubled the load found in the control snowpack (Table I). A substantial portion of the particulate matter found in both control and dust addition snowpacks was forest litter derived from the canopy (Table I). A similar proportion of the total particulate loading of the clearing snowpack in 2012 was also forest litter (Table I).

Our experimental treatment resulted in small differences in measured SWE and snowpack ablation between the control and dust addition treatments (Figures 3a–3c). There was significantly less SWE in the dust addition treatment when compared to the control during the accumulation period of 2010 ($p < 0.05$; Figure 3a). During 2011 and 2012, however, treatment differences in SWE accumulation were statistically indistinguishable (Figures 3b and 3c).

Table I. Snowpack particulate loading for 2011 and 2012 in a nearby clearing (no canopy) and in control and dust addition treatments (with canopy).[a]

	Clearing		Control		+ Dust	
	Total (SE)	Litter (SE)	Total (SE)	Litter (SE)	Total (SE)	Litter (SE)
2011	18.3 (NA)	NA	32.7 (8.5)	10.5 (5.1)	64.2 (20.2)	15.8 (8.0)
2012	19.1 (6.8)	13.3 (7.5)	38.5 (2.0)	24.9 (4.3)	73.4 (11.6)	49.8 (13.1)

[a] Mean total loading in g/m^2 and standard errors are given, along with the mean forest litter (in g/m^2) extracted from the total. Three full snowpack core samples were taken for each location/year, except in 2011 (explained in the text).

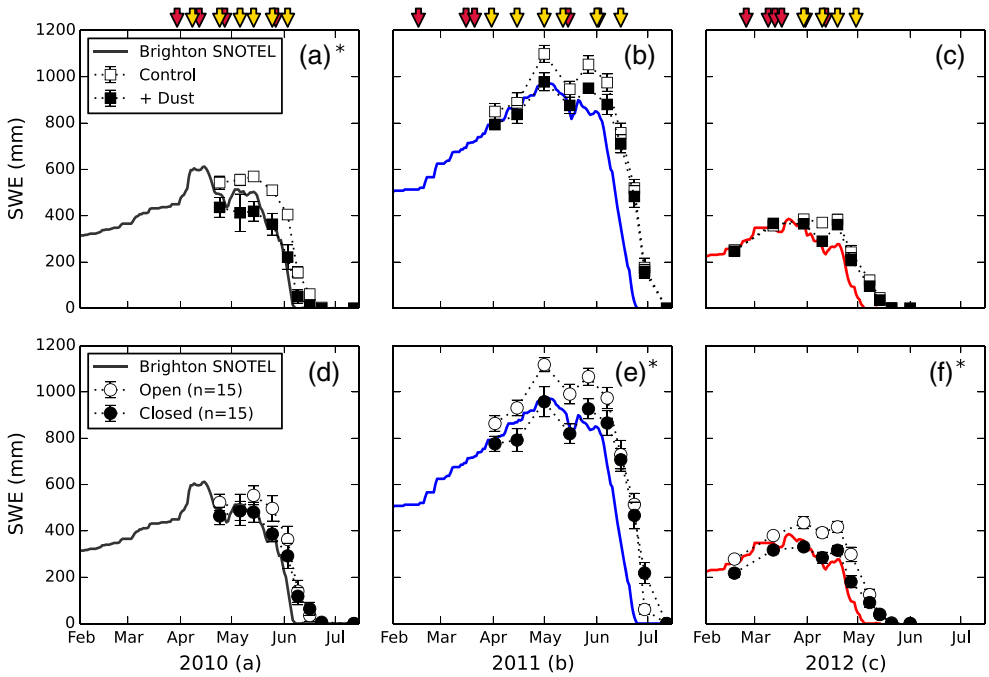

Figure 3. Snow water equivalent at the study plots. Observations were grouped and averaged in two ways in this figure. Panels (a), (b), and (c) compare the means of measurements made in control versus dust addition treatment plots ($n = 18$ per treatment). Panels (d), (e), and (f) compare the means of measurements made in open versus closed canopy locations in the forest ($n = 15$ per group, six median locations were excluded). Red arrows indicate the timing of natural dust events, and yellow arrows indicate experimental dust additions. Asterisks denote significant differences ($p < 0.05$) between control and dust addition SWE accumulation prior to the start of snowmelt (first four observations in each year). Letters in the x-axis labels denote significant post-hoc differences in SWE accumulation between years ($p < 0.05$). SWE observations from the Brighton SNOTEL site are shown for reference (as in Figure 2).

There was a large range of variability in canopy structure in our forest, and this influenced snow accumulation and ablation. Canopy openness, the percentage of a 180° sky view not occupied by the tree canopy, ranged from 16.7 to 50.7%. Canopy transmission, the percentage of above-canopy solar radiation (adjusted for seasonal solar zenith) transmitted to the forest floor, ranged from 11.5 to 68%. SWE was higher under open and high transmission canopy areas when compared to closed and low transmission canopy areas (Figures 3d–3f, canopy transmission groups not shown). Differences between open and closed canopies were statistically significant in 2011 and 2012 ($p < 0.05$), and differences between high and low transmission groups were significant in 2012 ($p < 0.05$).

Interannual variations in SWE at our study forest were much larger than the differences between treatments or between canopy groups in any single year. The mean of all SWE measurements (control and dust addition plots together) in the forest during the 2011 accumulation period was 898 mm (*S.D.* = 158 mm, $n = 36$), which was higher than the mean in 2010 (479 mm, *S.D.* = 120, $n = 20$) and 2012 (322 mm, *S.D.* = 79, $n = 36$). Pairwise post-hoc comparisons of accumulation period SWE between individual years indicated significant differences in SWE ($p < 0.05$; Figure 3).

Empirical ablation model

Visual inspection of the spring SWE depletion curves in 2010–2012 revealed similar ablation rates for control and dust addition treatments (Figures 3a–3c) but indicated a slightly higher rate in the open compared to the closed canopy groups (Figures 3d–3f). We tested whether this difference was significant by fitting a statistical model of snowpack ablation to our SWE measurements for 2011 and 2012 (Figure 4). Without interaction effects, our model fits the data reasonably well in 2011 ($R^2 = 0.70$) and 2012 ($R^2 = 0.78$). Air temperature, snowfall, and incoming solar radiation were all significant predictors of variation in *dSWE* in both years ($p < 0.002$), although given the structure of our model, their coefficients cannot be accurately compared.

Dust treatment and canopy structure both significantly impacted the ablation during at least part of the experiment. We tested several interaction terms in our statistical model to test whether the differences between treatment and canopy structure groups were significant. There were significant differences in ablation between the control and dust addition treatments in 2011 ($p < 0.02$). In 2012, however, the treatments were not statistically distinguishable. We also assigned each measurement location to an open or closed canopy and a high or low canopy transmissivity group and tested these groups as interactions in the model. Areas with high canopy transmissivity had faster ablation in 2011 ($p < 0.05$) and in 2012 ($p < 0.001$). Areas beneath an open canopy had faster ablation in 2011 ($p < 0.02$) and in 2012 ($p < 0.001$).

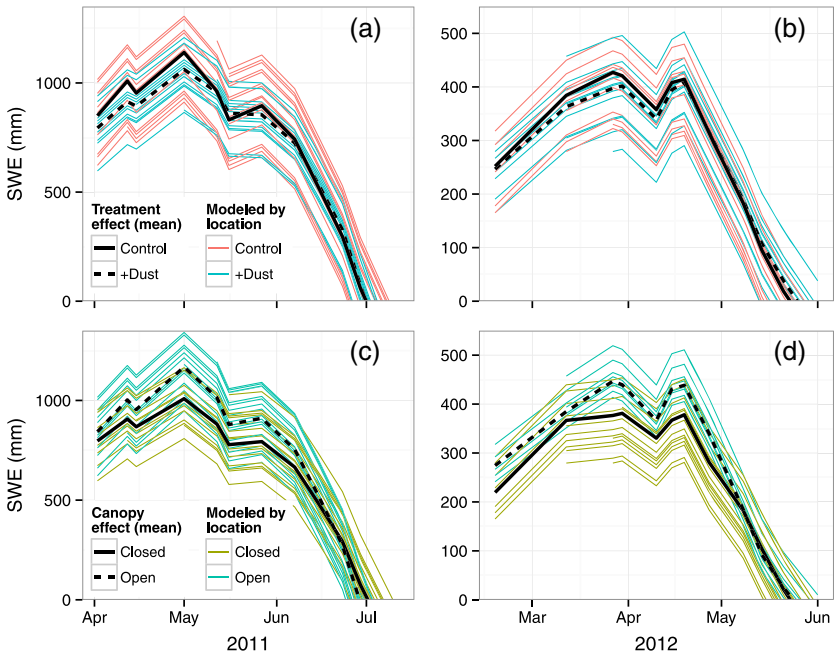

Figure 4. Modelled SWE during the spring of 2011 in panels (a) and (c), and the spring of 2012 in panels (b) and (d). Modelled SWE values were based on measured SWE during at the start of the experiment and linear model estimates of *dSWE* fit using measured SWE and climate data from the site. The model start day was different in 2011 and 2012, and two variations of the model were tested in each year. The model used in panels (a) and (b) includes a treatment interaction effect and in the panels (c) and (d) includes a canopy transmission interaction effect. Thick black lines represent mean SWE of all locations in each treatment or canopy group, beginning at each group's mean SWE on the starting day. Finer coloured lines are modelled for individual locations, beginning at each location's measured SWE on the starting day.

Soil temperature and water content

Average θ and T_{soil} were similar between treatments and canopy groups during 2011 and 2012. We constructed 95% confidence intervals around the mean T_{soil} and θ data from all sensors in control or dust addition plots and from all sensors classified as open and closed canopy (Figures 5 and 6). During the majority of each year, these intervals overlapped, indicating that the means of T_{soil} and θ were not statistically different between treatment or canopy groups. There were some minor differences in the dynamics of these variables between treatments or canopy groups that are detailed in the Discussion.

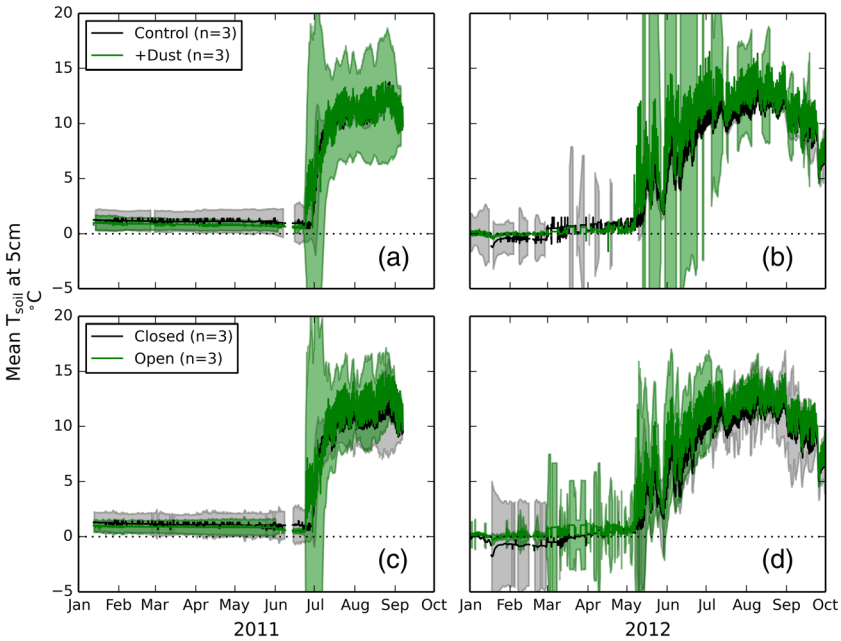

Figure 5. Comparison of mean soil temperature at 5 cm depth in the study plots. Lines represent the mean value of all sensors grouped by treatment in panels (a) and (b) or by canopy openness in panels (c) and (d). Shading represents the 95% confidence interval for data from all sensors used to calculate each mean. A dotted line at 0 °C is plotted for reference.

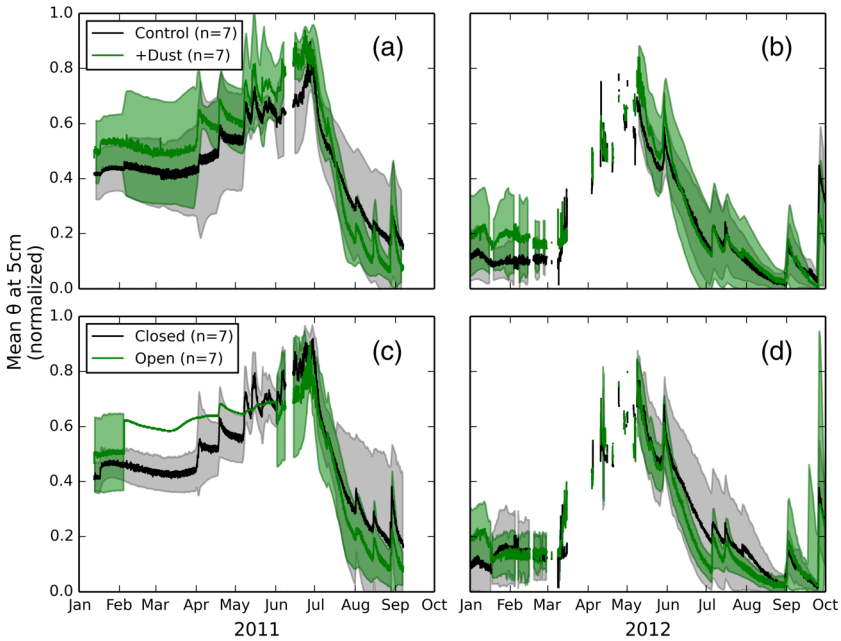

Figure 6. Comparison of mean volumetric soil water content (θ, normalized) at 5 cm depth in the study plots. Lines represent the mean value of all sensors grouped by treatment in panels (a) and (b) or by canopy openness in panels (c) and (d). Shading represents the 95% confidence interval for data from all sensors used to calculate each mean.

During the time periods when below-snowpack (January–June 2011 and 2012) and warm season (June–October 2011 and 2012) R_s measurements were made, we examined 24-h average values of T_{soil} and θ from soil profile sensors and handheld measurements. We found some significant differences in T_{soil} and θ between treatment and canopy groups that are presented in APPENDIX A and summarized in the Discussion. These differences were generally small and were associated with only minor variations in ecosystem processes amongst these groups.

Overall, interannual variability was the largest driver of differences in T_{soil} and θ (Figures A.11-A.16). Comparisons between years showed that below-snowpack, T_{soil} and θ were significantly higher in 2011 (large snowpack) than 2012 (small snowpack; $p < 0.05$). During the warm season, between-year θ comparisons indicated significantly wetter soils in 2011 than in other years ($p < 0.05$). This was observed with soil profile sensors at all depths and with the handheld sensors (2011 and 2012 only). Soils were generally warmest, measured by profile or handheld sensors, in 2012, but these differences were not significant.

Ecosystem processes

Ecosystem processes showed few significant differences between control and dust addition treatments in any year. Below-snow R_s was slightly higher in dust addition treatments compared to the controls in 2011 and 2012, but these differences were not statistically significant ($p > 0.05$; Figure 7). During the warm season there were no significant differences in R_s between treatments in any year (Figure 8). Xylem ψ did not vary in response to the dust treatment (Figure 9). Neither saplings nor mature firs showed significant differences in predawn or midday xylem ψ between control and dust addition treatments in any year tested.

The two pool decay models fit our litter bag mass loss data well, and there were small differences in litter decomposition rate between the treatments (Figure 10). The λ_f and λ_s for the control locations were 6.0×10^{-3} and 6.7×10^{-5}, respectively, and 7.8×10^{-3} and 8.2×10^{-5} for the dust addition treatment, respectively. The proportion of litter mass in the slow-cycling pool was slightly higher in the dust addition treatment (82% vs 77%), and the dust addition bags lost slightly less mass over the first winter.

There were significant differences in ecosystem processes between years in response to the widely varying winters. Below-snow R_s was significantly higher in 2011 than in 2012 (Figure 7; $p < 0.05$). Warm season R_s was significantly lower in 2010 than in the two following years

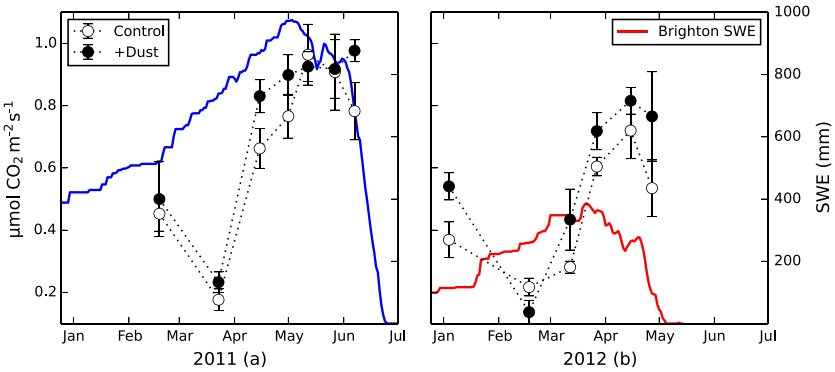

Figure 7. Mean values and standard errors of below-snowpack soil respiration measured in control and dust addition treatment plots ($n = 9$ for each treatment, left axis) during sampling dates in 2011 and 2012. Letters in the x-axis labels denote significant post-hoc differences in below-snowpack R_s between years ($p < 0.05$). There were no significant treatment differences. The Brighton SNOTEL SWE observations during the corresponding time period are shown with coloured lines for reference (right axis).

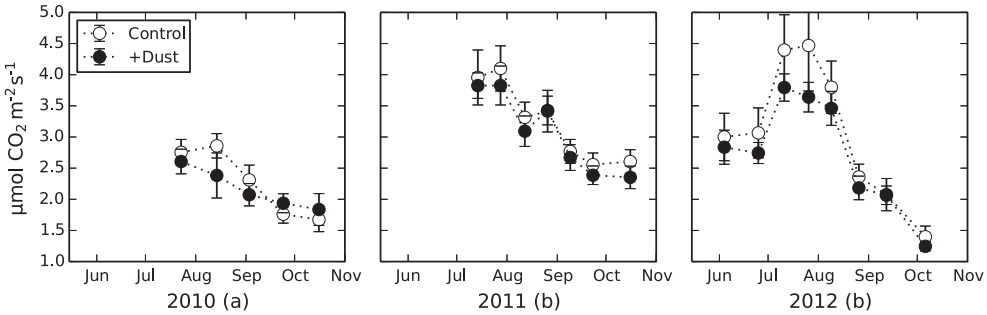

Figure 8. Mean values and standard errors of warm season soil respiration measured in control and dust addition treatments ($n = 18$ for each treatment). Letters in the x-axis labels denote significant post-hoc differences in warm season R_s between years ($p < 0.05$). There were no significant treatment differences. Note the difference in scale with Figure 7.

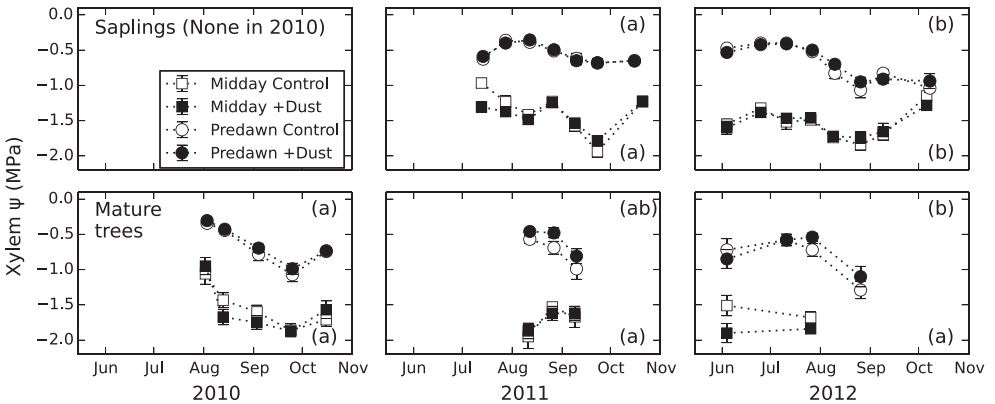

Figure 9. Mean values of xylem water potential in juvenile and mature subalpine fir measured in control and dust addition treatments ($n = 9$ for each treatment). No saplings were measured in 2010. Means and standard error bars, which are smaller than the symbols in many cases, are shown. Letters in the right corners of each panel indicate significant post-hoc differences in xylem ψ between years ($p < 0.05$). Top corner letters correspond to predawn ψ and lower corner letters to midday ψ. There were no significant treatment differences.

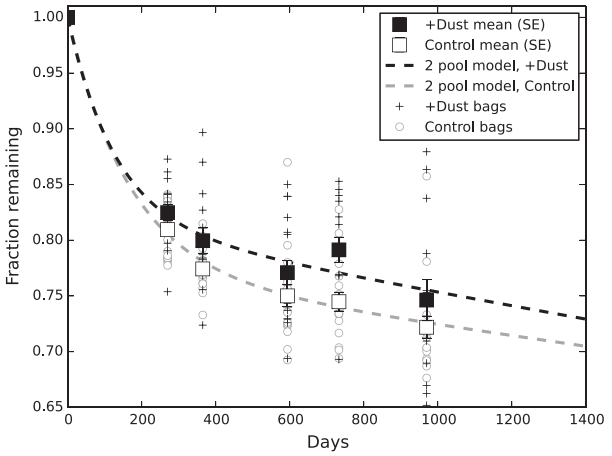

Figure 10. Litter bag mass loss between fall 2010 and spring 2013. Individual litter bag samples are shown (small circles or crosses), along with the mean and standard error of control and dust addition treatments for each collection date ($n = 18$ per treatment per date). The dashed lines were calculated using a two-pool exponential decay function fit using nonlinear least squares. Decay constants for each pool (λ_f and λ_s) in each treatment are given in the text.

($p < 0.05$), but respiration rates in 2011 and 2012 were not significantly different from each other (Figure 8). It is important to note that warm season respiration was measured at different and fewer locations in 2010. Sapling predawn ψ was lower in 2012 as was sapling midday ψ ($p < 0.05$). Water potential values of mature firs did not differ significantly between years (Figure 9).

DISCUSSION

Snow accumulation and melt

There were few statistically significant differences in SWE accumulation or ablation between the control and dust addition treatments. The primary radiative effect of dust or other particulate matter in snow is to lower the shortwave albedo of the snow surface and thereby increase its absorption of solar radiation. Secondary effects of dust, such as increases in snow grain size, exposure of below-snow surfaces, and changes in surface roughness also impact snowpack energy balance during the ablation season (Hansen and Nazarenko, 2004; Fassnacht *et al.*, 2009). Our dust addition treatment likely altered the energy balance of the snowpack by one or more of these mechanisms. Several possible reasons may explain the smaller than expected differences in ablation following dust addition. The first possibility is that the added dust did not significantly change the energy balance of the snowpack relative to the control. Another possibility is that added dust had a smaller effect on snowpack energy balance than did variations in snowpack energy balance resulting from differences in canopy structure within our forest. A third possibility is that higher accumulation and/or sublimation rates at open canopy locations in our forest compensated for faster ablation in the dust addition treatment. These three explanations are discussed below.

Our snowpack manipulation increased the dust load relative to the control but may have had a smaller than expected effect on albedo. When we measured the mass of particulate matter in our snowpacks near the close of each ablation season, the mass in the dust addition treatment exceeded the control by a factor of 2 (Table I). Given limited historical and projected data on dust emission and deposition in the western U.S., this seems like a plausible, and possibly conservative, dust loading rate for the coming century (Lawrence and Neff, 2009; Munson *et al.*, 2011). The control snowpack, however, had roughly double the particulate matter found in a nearby clearing. A large percentage of the total particulate matter in the control and dust addition treatments was composed of forest litter in 2011 (24–32% litter) and 2012 (64–70% litter), indicating

that particles other than our added dust probably impacted the snow surface albedo in both treatments, particularly in 2012. We do not have an explanation for the increase in forest litter in the snowpack in 2012. Snowpack albedo is often lower in forests when compared to clearings (Melloh *et al*., 2002), and a number of prior studies have indicated that forest litter is highly effective at reducing the albedo and increasing the ablation of subcanopy snowpacks (Hardy *et al*., 2000; Melloh *et al*., 2001; Winkler *et al*., 2010; Pugh and Small, 2012). Unfortunately, neither the spectral characteristics of dust and litter nor the snowpack albedo was quantified in this study, so it is difficult to judge the relative importance of dust versus forest litter.

Significant variation in ablation was explained by canopy structure and its effect on radiative energy balance. In alpine or other snowpacks without overstory vegetation, net shortwave radiation is commonly the most significant component of snowpack radiative energy balance during the spring ablation season (Marks and Dozier, 1992). In forested areas, incoming shortwave radiation is intercepted by the canopy (shading), and a portion of this absorbed energy is reemitted down to the snowpack as thermal radiation (longwave irradiance). The relative importance of shading and longwave irradiance to subcanopy snowpack energy balance depends greatly on canopy structure, temperature, and solar angle. In forests with open or discontinuous canopies, such as an aspen stand, there is less longwave irradiance to the snowpack from the canopy but greater transmission of shortwave radiation through the canopy. This becomes especially important in late spring as sun angle increases (Pomeroy and Dion, 1996; Hardy *et al*., 2004; Pomeroy *et al*., 2008; Lawler and Link, 2011; Lundquist *et al*., 2013). Accordingly, we found faster ablation under open canopy locations in our forest during the spring melt (Figure 4). As canopy closure increases, longwave irradiance also increases, and under some conditions, higher canopy longwave irradiance compensates for declines in shortwave transmission and becomes the major contributor of snowpack ablation energy (Link and Marks, 1999a; Sicart *et al*., 2004). In warmer climates, this may even lead to significant midwinter melt events (Lundquist *et al*., 2013). We did not observe such midwinter melt events, but it is possible that longwave irradiance was the dominant contributor to ablation energy in closed canopy locations prior to and during the spring ablation period. The albedo effect of dust acts primarily in the solar portion of the spectrum. If the largest portion of the energy available for snowpack ablation in this forest was canopy longwave irradiance, the dust addition treatment would have been less effective in perturbing snowpack energy balance.

The potential for snow interception, sublimation, and redistribution also varied with canopy structure in our forest. Forest canopies intercept snowfall and facilitate water loss through redistribution and sublimation (Hedstrom and Pomeroy, 1998). Consequently, it is common to find greater snow accumulation beneath forest canopy openings relative to closed canopies (Hardy *et al*., 1997; Koivusalo and Kokkonen, 2002). Our data clearly showed that more SWE accumulated beneath an open canopy (Figures 3d–3f), indicating less snowfall interception, sublimation, and/or redistribution in these areas. If we assume that dust addition lowered snowpack albedo and thus hastened ablation in our study forest, the effect would be highest in these same open areas where greater shortwave radiation was available to melt snow. It is possible that higher accumulation rates compensated for faster ablation in dust addition locations, making differences between treatments difficult to observe. Similar compensatory effects have occurred in other forest snowpack studies. Biederman *et al*. (2012) found lower snow interception during the grey phase in a mountain pine beetle impacted forest stand (presumably more open), but this was compensated for by higher sublimation rates in these stands.

Our empirical model results support the idea that canopy structure had a similar or greater effect on snowpack ablation and accumulation than dust. Although our empirical model was not a full energy balance model, it successfully reproduced changes in SWE in our study forest. Snow accumulation was slightly higher in the control plots than in dust addition plots during the accumulation phase of each year (Figure 4). Given that, on average, control locations had a slightly more open canopy than the dust addition treatment (Table II), it is unclear whether this occurred because of the effects of dust or canopy structure. Later in the spring of 2011 and 2012, the control and dust addition treatments showed similar rates of ablation, indicating that dust had a small effect on snowpack energy balance between treatments (Figure 4). In both years, however, there were significant differences in snowpack ablation below high and low transmission canopies, indicating that differences in canopy structure led to differences in snowpack energy balance. Snowpacks below more open canopies also had significantly greater snow accumulation during early spring, probably due to low canopy interception. These results agree with numerous other studies demonstrating greater accumulation and more rapid snow ablation beneath openings in forest canopies (Hardy *et al*., 1997; Koivusalo and Kokkonen, 2002; see Varhola *et al*., 2010 for a review; Musselman *et al*., 2012a). Together, these two effects resulted in similar snowpack melt out timing below these contrasting canopy types (Figure 4). It should be noted, however, that in warmer forests or during warm years, significant winter melt events may occur beneath dense canopy cover because of longwave irradiance. Under these conditions,

Table II. Means of the canopy structure measurements derived from hemispherical photographs, including the percentage of sky view not occupied by canopy (% Open) and the percentage of incoming solar radiation transmitted by the canopy (% Transmitted) in control and dust addition treatments.[a]

		Control		+Dust	
	Meas. location	% Open	% Transmitted	% Open	% Transmitted
With leaves	Litterbags	19.4 (5.9)	30.8 (11.7)	21.8 (4.4)	34.8 (11.2)
	Soil respiration	19.6 (6.4)	31.8 (12.5)	20.9 (4.4)	32.5 (11.0)
	Soil profiles	19.6 (4.9)	33.1 (10.1)	22.5 (3.9)	34.9 (8.4)
Without leaves	SWE locations	30.2 (8.8)	45.0 (14.2)	27.7 (5.9)	40.6 (12.6)
	Soil profiles	22.9 (4.5)	34.1 (13.0)	24.4 (7.7)	35.7 (17.3)

[a] Locations were photographed during time periods with overstory deciduous leaves present, without deciduous leaves present, or both. Standard deviations of the means are in parentheses.

seasonal snowpacks may persist longer in open canopy areas (Lundquist *et al*., 2013).

Our results indicate a high dependence of snow accumulation and ablation on canopy structure and highlight the need for more detailed study of subcanopy dust-on-snow effects. Although this is the first such dust manipulation in a forested area, scientists have applied distributed hydrological models to calculate the effect of dust deposition on snowpack dynamics and spring runoff across large areas of the western U.S. (Painter *et al*., 2007, 2010). Models used in these studies employ realistic, full energy balance calculations for forested areas, but the driving data for overstory vegetation, subcanopy albedo, and their effects on snowpack energy balance tend to be coarsely defined. The VIC model, for example, uses a 1 km vegetation grid, with leaf area index (LAI) specified for the vegetated fraction of each grid cell using a global LAI database derived from 1981–1994 averages values (Liang *et al*., 1994; Myneni *et al*., 1997; Gao *et al*., 2010). Solar radiation attenuation, longwave irradiance, snow interception and redistribution, and other canopy-dependent snowpack energy and mass balance parameters are calculated on the basis of this gridded data. With realistic estimates of subcanopy solar and thermal radiation, accurate estimates of snowpack dynamics can be made at point or distributed scales (Link and Marks, 1999a, 1999b; Musselman *et al*., 2012b), but obtaining or estimating these data at or beyond the watershed scale is not an easy task. Our results suggest that under open, heterogeneous canopy cover, which is common in western U.S. mountains, forest canopy has an effect on snowpack ablation that is slightly larger in magnitude than the effect of dust.

The shortwave albedo of subcanopy snowpack has an underappreciated role in determining snowpack radiative energy balance. A sensitivity study by Sicart *et al*. (2004) found that when subcanopy snow albedo was high, the radiative energy balance of the snowpack changed little in response to variation in canopy density. At low albedo (<0.5), however, the radiative energy balance of the snowpack was sensitive to increases in shortwave transmission through a canopy. Thus, aeolian dust deposition should be expected to alter the radiative energy balance of some forests. A number of studies provide interesting context, but many of these have taken place in disturbed forests. In the western U.S., where the mountain pine beetle is currently impacting forests at a large scale, Pugh and Small (2012) found that high rates of litter deposition in beetle impacted conifer forests lowered snowpack albedo. They estimated that this increased snowpack ablation to a greater extent than other radiative or atmospheric effects resulting from tree death in the forest. Gleason *et al*. (2013) found a 200% increase in net shortwave radiation at the snowpack surface in a recently burned conifer forest. This change was due to the combined effects of higher solar radiation transmission by the canopy and lower snowpack albedo due to the deposition of burned woody debris. So, although it is established that changes in albedo impact the energy balance of a subcanopy snowpack, the conditions under which this results in faster ablation are not documented in a broad number of forest types, with notably few studies in undisturbed forests such as ours. Without more detailed, spatially explicit data on canopy structure and subcanopy snowpack albedo, it remains challenging to predict the effect of aeolian dust deposition on subcanopy snowpacks at a large spatial scale.

Impacts on the soil biophysical environment

Our snowpack manipulation had few effects on the soil environment. Differences in the overall seasonal pattern of mean T_{soil} and θ in the control and dust addition treatments were not significant (Figures 5 and 6, panels a and b), which is consistent with the small effects our treatment had on snowpack dynamics during the spring. Differences between the seasonal pattern of mean T_{soil} and θ in open and closed canopy groups were also insignificant (Figures 5 and 6, panels c and d). Despite this, there was an interesting difference in seasonal T_{soil} patterns between

treatment and canopy groups. Surface T_{soil} began to increase from a near-zero level below the snowpack at or near the time snowcover disappeared, consistent with other observations in snow-covered ecosystems (Lundquist and Lott, 2008). This occurred a few days earlier in the dust addition plots (compared to the control) and open canopy locations (compared to closed) during 2011 and 2012 (Figure 5), perhaps indicating an earlier completion of ablation, on average, in these groups. Open canopy areas had greater radiative exposure that may have led to greater evapotranspiration (Molotch *et al*., 2009; Bales *et al*., 2011) and earlier decline in surface θ during the spring (Figures 6c and 6d).

When we examined T_{soil} and θ at discrete time periods (concomitant with below-snowpack or warm season R_s measurements; see APPENDIX A for details) there were some significant differences between treatment and canopy groups. Dust addition plots and open canopy locations had wetter soil and lower temperature below the snowpack in some years (Figures A.3 and A.4), perhaps indicating that lower snowpack albedo or greater radiation transmission led to more frequent winter melt events that delivered cold melt water to the soil profile. We view this as somewhat unlikely in our high-elevation forest, although such events might be common where snowpacks are at or near an isothermal state during winter (Bales *et al*., 2011). In the warm season, T_{soil} beneath open canopy was higher than beneath closed canopy (2012 only, Figure A.2), again suggesting greater radiation exposure in these areas. There were few consistent differences in warm season θ between treatment or canopy groups.

Interannual variability in T_{soil} and θ was larger than any difference due to dust treatment or canopy structure. The large snowpack year, 2011, had the highest below-snowpack T_{soil} (Figure A.1), indicating that the large snowpack effectively insulated the soil from the temperature and radiative environment at the snow surface (Zhang, 2005; Maurer and Bowling, 2014). Below-snowpack and warm season θ were also higher in 2011 (Figures A.3 and A.6), suggesting that there was greater infiltration of snowmelt water into the soil in this year. Warm season θ was lowest during 2012, the year with the smallest snowpack. These patterns held at most soil depths.

Impacts on ecosystem processes

In alpine areas, earlier snowmelt has a marked effect on plant and animal phenology (Inouye *et al*., 2000; Steltzer *et al*., 2009), and on the basis of present understanding of ecosystem processes we anticipated similar impacts to carbon and water cycling in our study. Differences in ecosystem carbon and water cycle processes between control and dust addition plots were not significant in the majority of cases (Figures 7–9). We attribute this to the small differences in T_{soil} and θ between these treatments, and had our snowpack manipulation accelerated snowpack ablation as we expected, perhaps carbon and water cycle processes would have responded. Litter decomposition rate was slightly slower in the dust addition treatment (Figure 10), but there were no consistent differences in T_{soil} or θ between the treatments that explained this.

There were significant differences in ecosystem processes between the years of our study, which varied greatly in snowpack size and melt timing. As expected, interannual variability in the soil environment was related to snowpack characteristics, and the resulting year-to-year differences in T_{soil} and θ (see APPENDIX A) appeared to drive variation in R_s and xylem ψ between years. Of the three winters observed in our experiment, soils were warmest and wettest below the 2011 snowpack, and the highest below-snow R_s occurred in this year. A number of studies have highlighted that significant amounts of CO_2 are respired from soil below seasonal snowpacks and that these fluxes may vary significantly in response to changes in the below-snowpack soil environment (Monson *et al*., 2006a, 2006b; Liptzin *et al*., 2009; Aanderud *et al*., 2013). Snow molds, for example, are a group of fungi that colonizes forest litter below Rocky Mountain (and probably other) snowpacks in the spring and are highly sensitive to small fluctuations in temperature (Schmidt *et al*., 2009). Soil microbial physiology such as this may explain the higher below-snowpack respiration rate we observed in 2011.

Of the warm seasons observed in our experiment, soils were driest following the small 2012 snowpack. This did not impact R_s but did influence water availability for trees. Low warm season θ in 2012 resulted in predawn and midday sapling ψ that was significantly lower than other years. This result agrees with other studies in our region indicating that years with lower SWE and earlier snow melt result in diminished soil water availability for vegetation (Molotch *et al*., 2009; Hu *et al*., 2010).

CONCLUSIONS

We artificially increased the load of aeolian dust in a subcanopy mountain snowpack in an effort to change snowpack albedo and radiative energy balance. This dust addition treatment had a relatively small impact on snow accumulation, snowpack ablation, and the timing of snowmelt in our study forest. The influence of the canopy, through the combined effects of snow interception and shading, produced a somewhat larger impact on snowpack accumulation and ablation. Both SWE amount and ablation were significantly greater beneath open as compared to closed canopy areas in our study forest. Very significant differences in snowpack size and melt timing

resulted from interannual snowfall variability, although given that 2 years of the study had near record high and low snowpacks, this variability cannot be considered the norm for this region.

Dust addition produced few significant effects on the soil environment or on ecosystem processes. There were, however, significant differences in ecosystem processes between years, and this interannual variability was larger than any within-year effect of dust or canopy. Interannual differences in soil temperature and soil water content were in the direction expected given the year's snowpack size and melt timing. The resulting variation in the soil environment appeared to drive the differences in ecosystem processes we observed.

The limited impact of our dust manipulation in this forest suggests that the effect of aeolian dust on snowpack ablation is additive with numerous other site specific energy and mass balance factors. In this system, within-forest and interannual variation in snowpack mass and energy balance were larger than the effect of dust, although only moderately so in the case of canopy structure. Both field and modelling studies of the influence of aeolian dust on snowpack ablation would benefit from better representation of canopy and its influence over snowpack energy balance. Future research on this topic should target interactions between canopy structure, climate variability, and snowpack albedo to better understand the conditions under which dust deposition may influence ecohydrological processes.

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APPENDIX A. TREATMENT AND CANOPY EFFECTS ON SOIL TEMPERATURE AND WATER CONTENT

In our examination of multisensor mean time series (Figures 5 and 6) we found few differences between dust addition treatment and canopy groups. We also examined T_{soil} and θ during the discrete time periods used for sampling R_s below the snowpack and in the warm season. To compare treatment, canopy, and interannual differences in below-snowpack T_{soil} and θ, we calculated 24-h average T_{soil} and θ values from soil profile sensor data on each below_snowpack R_s sampling date (January–June 2011 and 2012). For comparison of warm season T_{soil} and θ, we calculated 24-h average T_{soil} and θ values from soil profile sensors and collected handheld T_{soil} and θ measurements for each warm season respiration measurement date (June–October 2010, 2011, and 2012). Handheld measurements were made at all R_s collars. T_{soil} was measured at 5 and 15 cm depths using a thermocouple soil probe, and θ measurements were integrated across the top 10 cm of the soil profile using a time–domain measurement soil water content probe (see Methods for instrumentation details). We tested for treatment and canopy differences in mean T_{soil} and θ data using a multilevel linear model with sample date as a random variable (a repeated measures design). To test for interannual differences in mean T_{soil} and θ, measurement location was the random variable because the same locations were measured in all years. This was followed by post-hoc analysis (Tukey's honest significant difference), and we report post-hoc test results ($\alpha = 0.05$) to indicate significant differences between the means of each year.

Soil differences between treatment and canopy groups

There were a number of significant differences in T_{soil} and θ between treatment and canopy groups during the below-snowpack period. In 2011, control plots were significantly warmer (higher T_{soil}) than dust addition plots ($p < 0.05$–0.0001, depending on depth), but we found no significant difference between treatments in 2012 (Figure A.1). There were no consistent differences in mean T_{soil} between the canopy groups, although open and closed canopy mean T_{soil} were significantly different in some years and at some depths ($p < 0.05$–0.01; Figure A.2). Dust addition plots had significantly wetter soil (higher mean θ) than control plots below the snowpack in 2011 ($p < 0.001$), but there were no significant θ differences in 2012 (Figure A.3). Open canopy locations were also significantly wetter at 5 and 60 cm depths below the snowpack during both years ($p < 0.05$–0.0001; Figure A.4).

During the warm season, T_{soil} and θ were also significantly different in some treatment and canopy group contrasts. In 2012, 5 cm soil sensors indicated warmer T_{soil} in the dust

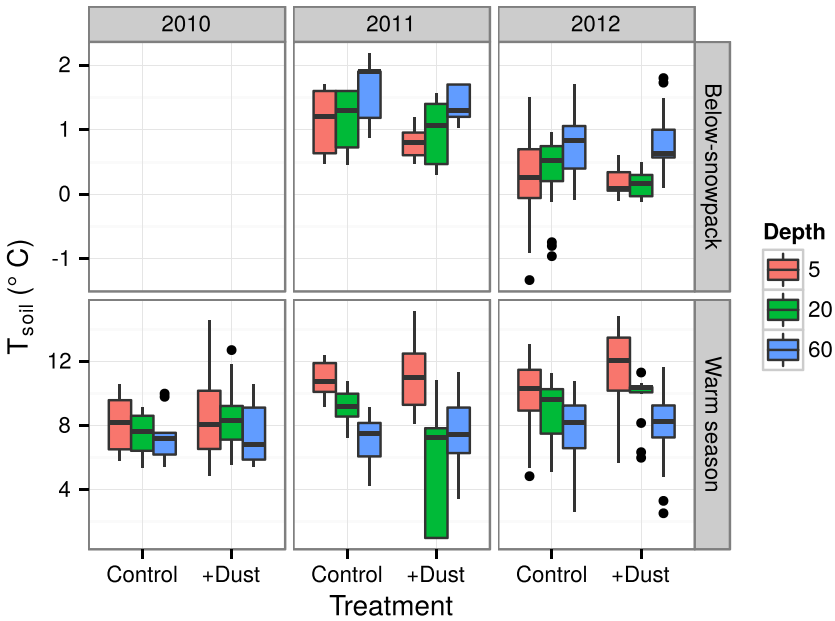

Figure A.1. Comparison of 24-h mean T_{soil} taken from soil sensor profile data collected during the below-snowpack (top panels) and warm season (lower panels) R_s measurement periods in 2010, 2011, and 2012. The three soil sensor depths are shown, and sensor means are split into + Dust and Control treatments. Boxplots show the range (bars), first and third quartile (top and bottom of box), and median (line within box) of the data in each group. There were no below-snowpack respiration measurements made in 2010.

addition plots ($p < 0.01$; Figure A.1), but there were no other significant treatment effects observed using soil profile sensors or handheld measurements (5 and 15 cm measurements; Figure A.5). Open canopy locations had significantly warmer soils in 2012 ($p < 0.05$–0.0001, depending on depth), but handheld T_{soil} measurements did not corroborate these differences (Figures A.2 and A.5). In 2012, deep soil profile sensors (20 and 60 cm) had higher θ in the dust addition plots compared to the controls ($p < 0.05$–0.01 depending on depth; Figure A.3). Mean 20 cm θ was slightly lower in open canopy locations in 2010, but otherwise there were no other canopy or treatment differences in warm season θ observed in profiles or handheld measurements (Figures A.4 and A.6).

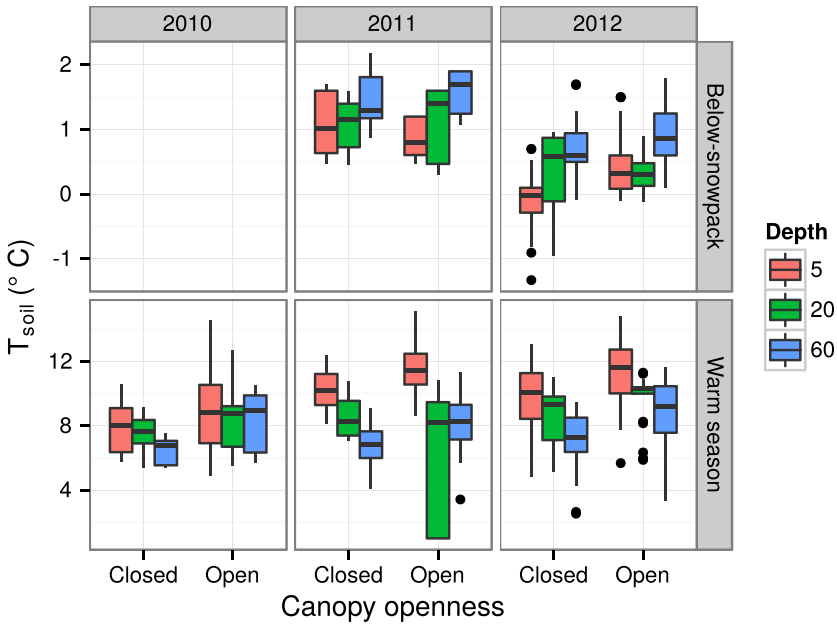

Figure A.2. Comparison of 24-h mean T_{soil} from soil sensor profiles as in Figure A.1. In this figure, sensor means are split into Open and Closed canopy structure groups.

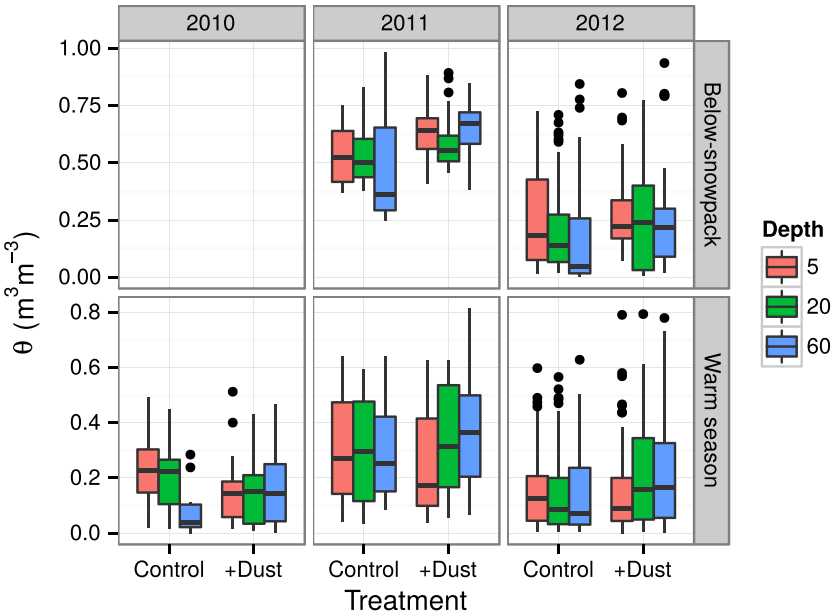

Figure A.3. Comparison of 24-h mean θ taken from soil sensor profiles as in Figures A.1 and A.2. In this figure sensor means are split into + Dust and Control treatments.

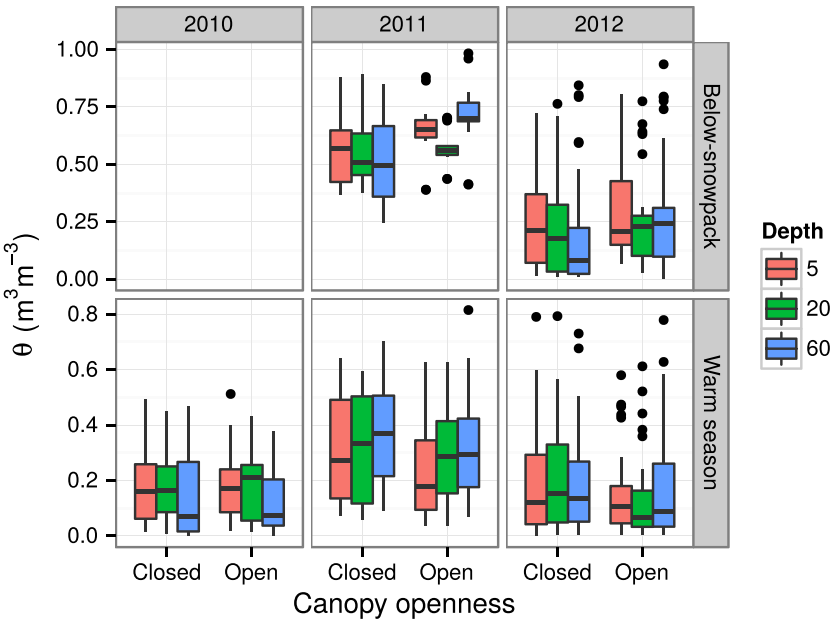

Figure A.4. Comparison of 24-h mean θ from soil sensor profiles as in Figures A.1, A.2, and A.3. In this figure, sensor means are split into Open and Closed canopy structure groups.

Soil differences between years

We found larger and more statistically significant differences in T_{soil} and θ between years, and it is likely that this was driven by the large interannual variation in snowpack size during our study (Figure 2). In the largest snowpack year, 2011, below-snowpack T_{soil} and θ were significantly higher than in other years ($p < 0.05$) and these patterns held at all depths (Figures A.1 and A.3). During the warm season, soils were wettest in 2011 (large snowpack year) and significantly drier in the lowest snowpack year, 2012 ($p < 0.05$), and this pattern held at all depths and was significant with soil profile and handheld probe data (Figures A.3 and A.6). There were not, however, significant differences in warm season T_{soil} between years, whether measured by soil profile sensors or the handheld probe (Figures A.1 and A.5).

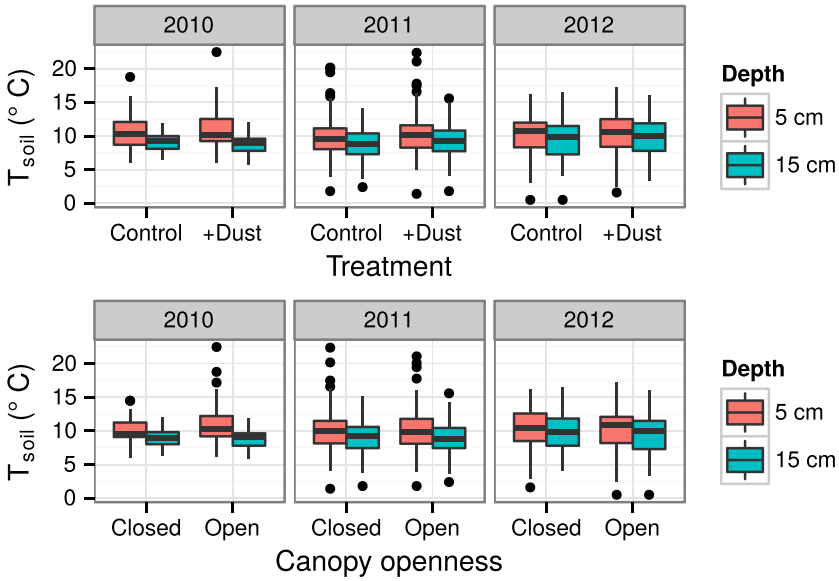

Figure A.5. Comparison of mean T_{soil} measurements made with a handheld probe during warm season R_s measurements (2010–2012). Means are split into + Dust and Control treatments in the top panels and Open and Closed canopy structure groups in the lower panels.

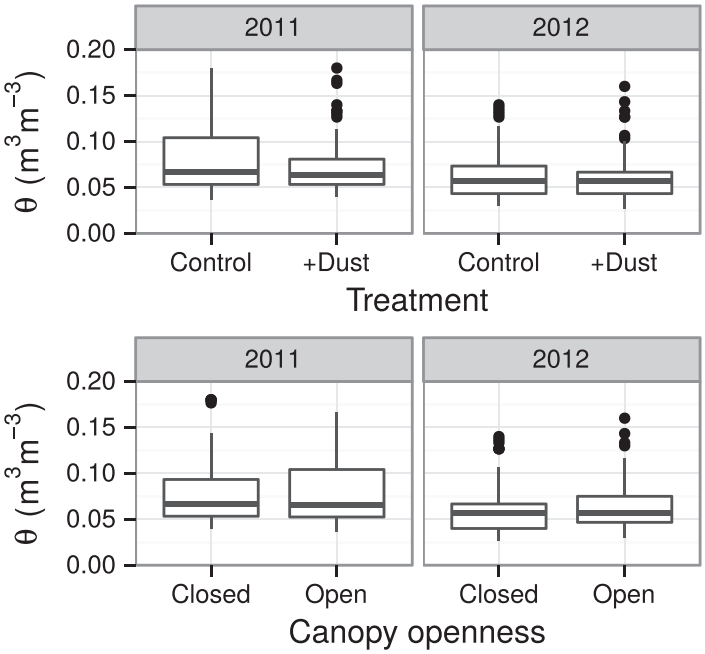

Figure A.6. Comparison of mean θ measurements made with a handheld probe during warm season R_s measurements (2011 and 2012). Means are split into groups as in Figure A.5.

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