



Using inexpensive temperature sensors to monitor the duration and heterogeneity of snow-covered areas

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[1] Small, self-recording temperature sensors can be deployed quickly and inexpensively to monitor spatial and temporal patterns of snow accumulation and melt in complex environments. Burying these sensors slightly below the soil surface provides a record of the presence or absence of snow cover because near-surface soil temperatures only experience diurnal temperature oscillations when they are not covered by an insulating layer of snow. When combined with an air temperature record and snowmelt model, the date snow cover disappears can be used to approximate the amount of snow that accumulated at the start of the melt season.

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1. Introduction

[2] Variability in snow-covered areas (SCA), i.e., snow patchiness, occurs at scales ranging from one meter to many kilometers, and an accurate depiction of this spatial heterogeneity of snow-covered area is necessary to accurately model the magnitude and timing of snowmelt runoff and land surface climatic feedbacks, such as albedo [Essery and Pomeroy, 2004; Giorgi and Avissar, 1997; Liston, 1999, 2004; Luce et al., 1998; Skaugen, 2007]. Here we present a brief overview of prior work on snow variability and present data from both the Sierra Nevada and Colorado Rockies to demonstrate how soil temperature measurements can be used to learn about the spatial distribution of snow.

2. Prior Work on Snow Variability

[3] Variability in SCA, snow depth, and snow water equivalent (SWE) are all closely related. Because density varies less than snow depth and is random in nature at the hillslope scale [Anderton et al., 2004; Shook and Gray, 1997], most of the spatial variability in SWE is due to variability in snow depth. In many locations, the initial snow depth distribution varies more spatially than melt rates [Dunn and Colohan, 1999; Hartman et al., 1999; Luce et al., 1998]. Because of this, the date of snow disappearance can be combined with a snowmelt model to back out the distribution of SWE when melt began [Liston, 1999; Lundquist et al., 2005; Molotch and Bales, 2005]. Thus, knowledge of spatial patterns in any one of snow depth, SWE, and snow disappearance date can generally be used to infer spatial patterns in the other parameters [Liston, 1999].

[4] The spatial and temporal evolution of SCA is generally monitored with satellite imagery, (e.g., Landsat, MODIS or AVHRR, with 30-m, 500-m, and 1000-m reso-

lutions, respectively), and with sparsely located surface measurements of snow depth and SWE. Fractional SCA at scales smaller than these observational scales can be computed using linear spectral unmixing of multispectral remotely sensed data [Rosenthal and Dozier, 1996; Painter et al., 2003], but is generally parameterized with a snow cover depletion scheme, many of which are summarized by Essery and Pomeroy [2004]. A proper parameterization requires knowledge of the local variance in SWE or snow depth, which changes with the scale of interest [Blöschl, 1999].

[5] Snow depth variability has been measured with intensive depth probe sampling [Elder et al., 1991], photographs [Tappeiner et al., 2001; König and Sturm, 1998], radar [Machguth et al., 2006], and lidar [Deems et al., 2006]. At the hillslope scale (1–100 m), snow depth varies primarily because of wind and avalanche redistribution [Deems et al., 2006; Elder et al., 1991; Anderton et al., 2004] and variations in vegetation (e.g., forested versus open areas [Pipes and Quick, 1987]). At hillslope scales and in exposed prairie and arctic environments, snow depth distributions are generally fractal at small scales and random at scales beyond some limiting length, which has been observed at lengths between 10 and 100 m [Deems et al., 2006; Shook and Gray, 1997; M. Clark et al., The use of field data to design distributed snow models, submitted to *Water Resources Research*, 2008]. At larger scales (>100 m), slope, elevation, net radiation, precipitation amount, and air temperature become more important [Erxleben et al., 2002]. In areas with a predominant winter wind direction, which includes the mountains of the western continental U.S., snow depth distribution patterns tend to repeat from one year to the next, so an intensive study carried out one winter can be used to spatially distribute limited snow depth measurements in future years [Deems et al., 2008; Erickson et al., 2005; Molotch and Bales, 2006].

3. Near-Surface Soil Temperature Under Snow

[6] Near-surface soil temperature measurements provide a robust and relatively inexpensive method to track the

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Table 1. Comparing Self-Recording Temperature Loggers With DTS

	Self-recording Temperature Loggers	DTS
Approximate unit cost in 2008	\$30 per sensor, stand alone	\$35,000 logger + \$0.50–1.70 per meter of fiber optic cable
Programming requirements	Basic PC, USB-based connector, each sensor programmed individually with simple user-interface	Dedicated logger or desktop computer with data acquisition and processing software [Selker <i>et al.</i> , 2006a, 2006b]
Deployment requirements	Each sensor deployed and collected individually each year. Some instruments must be wrapped with plastic to prevent corrosion. To back out SWE, need additional sensor in the air, can be placed in a tree [Lundquist and Huggett, 2008]	Need instrument shelter and data logger for DTS, some kind of protection for cable (to protect from rodents and other hazards). Cable length of 100 m to 30,000 m can be deployed with any orientation.
Accuracy	$\pm 0.2^{\circ}\text{C}$	$\pm 0.1^{\circ}\text{C}$
Resolution/precision	For long deployments 0.5°C	0.01°C
Sampling frequency	1 s to 24 h (set at 1 h for 11-month deployment)	10–60 s
Spatial resolution	The spacing between sensors (could be as small as 10 cm)	1 m
Data retrieval	Instruments must be manually retrieved, generally done annually	Real-time delivery possible
Most practical for four deployment scenarios	<ol style="list-style-type: none"> 1. Citizen-based science (easy to deploy and retrieve) 2. Only a few measurement points are needed 3. Measurement locations are quite distant from each other, with intervening stretches that would be impractical for cables to traverse 4. Wilderness locations where unobtrusive measurements are needed 	<ol style="list-style-type: none"> 1. Dense sampling on a hillslope is required 2. Large number of measurements required where cable can be stretched between them 3. Real-time data are needed 4. For applications where the subgrid-scale fractal structure of snow variability needs to be resolved

distribution of snow cover. Snow has a thermal conductivity 5 to 20 times lower than that of mineral soils, ranging from 0.10 to $0.50 \text{ W m}^{-1} \text{ K}^{-1}$ [Zhang, 2005]. Thus, the presence of snow cover critically affects soil temperatures and ground heat fluxes [Sokratov and Barry, 2002; Taras *et al.*, 2002; Zhang, 2005].

[7] In arctic and alpine regions, with extreme cold temperatures and often shallow (<30 cm) snowpacks, near-surface soil temperatures track air temperatures, but with a damped signal [Taras *et al.*, 2002]. In these extreme environments, the beginning of snowmelt can be identified by the time when the snow-soil interface temperature rises to 0°C and remains there for several days, indicating melting snow surrounding the thermistor [Taras *et al.*, 2002]. In more temperate regions, below tree line, and where snowpacks are deep (>50 cm), near-surface soil temperatures are consistently 0°C so long as snow cover is present [Tyler *et al.*, 2008].

[8] Thus, temperature sensors placed 2–10 cm beneath the soil surface provide not only an indication of soil temperatures and frozen ground but, when distributed, also provide spatial representations of the presence or absence of snow cover. Tyler *et al.* [2008] used fiber-optic distributed temperature sensing (DTS) technology to obtain basal snow temperatures at resolutions of 1 m, 10 s, and 0.1°C in both California and Idaho. They observed uniform basal snow temperatures of 0°C and highly varying diurnal ground temperatures in snow-free areas. While fiber-optic cable costs as little as 50 cents per meter, the entire setup, including DTS technology, costs over \$30,000 (Table 1). Below, we present an alternate measurement method.

4. Temperature-Sensed Snow Cover With Distributed Self-Recording Sensors

[9] Self-recording temperature sensors, such as Maxim iButtons [Hubbart *et al.*, 2005] and Onset Tidbits, Pendants,

and HOBOs [Whiteman *et al.*, 2000], with prices as low as \$30 per sensor, have an accuracy of better than $\pm 0.5^{\circ}\text{C}$ and an ability to record hourly data for a duration of about a year or more (Table 1). Many of these sensors are waterproof and can be submerged in lakes and streams (e.g., Onset Tidbits), while others are water resistant and must be protected from corrosion if placed in contact with water for long periods of time (e.g., Maxim iButtons, Figure 1).

[10] A set of soil temperature sensors (Maxim iButtons) were distributed in Yosemite National Park, California (Figure 2) and on Niwot Ridge, Colorado (Figure 3) from September 2005 to June 2006. Sensors were wrapped in thin plastic (Figure 1) to prevent corrosion, buried 2 to 20 cm beneath the surface (mainly for protection from rodents), and linked with a nylon cord to a nearby tree trunk, root, or marker to aid in finding the sensor again. Sensors were buried at least 0.5 m from tree trunks to avoid monitoring the “tree well effect,” where snow immediately adjacent to the trunk melts first in the spring because of enhanced



Figure 1. An iButton (Maxim DS1922L) wrapped in plastic wrap, ready to be buried in the ground below a tree.

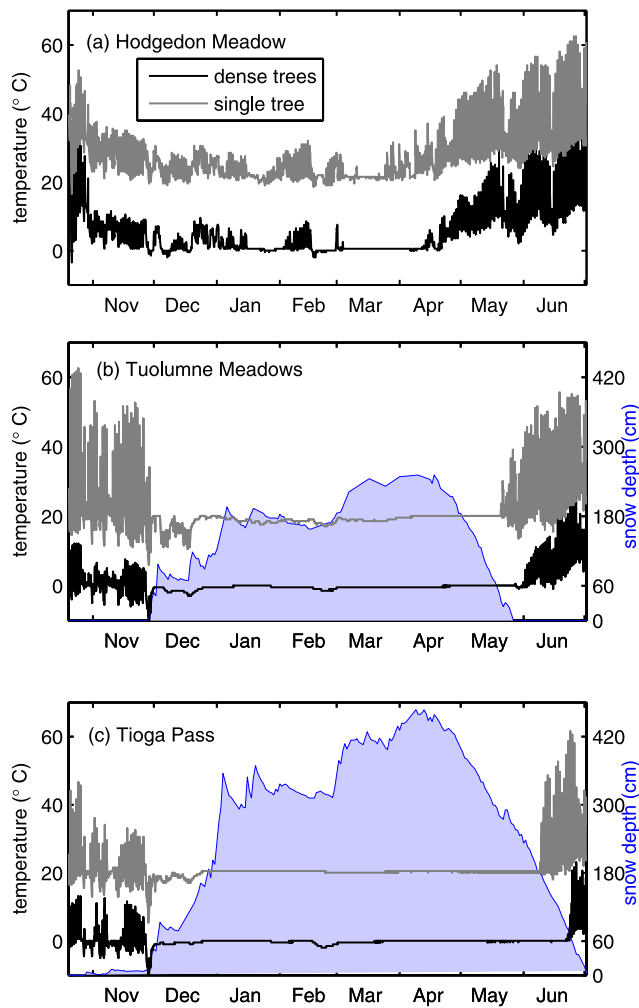


Figure 2. Hourly temperature data (left axis) from sensors deployed 5 cm below the soil surface at (a) Hodgedon Meadow, 1500 m elevation, (b) Tuolumne Meadows, 2600 m, and (c) Tioga Pass, 3000 m, below a single tree (with no other trees within 3 m) and a dense group of trees (continuous forest cover, with trees touching one other) for September 2005 to June 2006. At Hodgedon Meadow, the dense trees were white fir (*Abies concolor*), and the single tree was a Ponderosa pine (*Pinus ponderosa*). All other trees were lodgepole pines (*Pinus contorta*). Also shown is snow depth (right axis) measured in a clearing less than 50 m away. The temperature record from the single tree at each site has been shifted up by 20°C.

longwave radiation. Sensor locations were identified with GPS coordinates, photographs, and drawings. The sensors and their data were retrieved the following summer.

[11] Figures 2 and 3 illustrate examples of how distributed temperature sensors can be used. In Yosemite National Park, ground-based temperature sensors were placed at the base of an isolated tree (with no other trees nearby within 3 m) and in stands with continuous forest cover (all trees touching at least one other) at multiple elevations. At Tuolumne Meadows and Tioga Pass, these records were compared with California Department of Water Resources (CA DWR) sonic depth sensor measurements of snow depth in a clearing less than 50 m from the trees. The

temperature sensors recorded a steady near-0°C temperature when they were covered with snow (Figure 2). Snow cover lasted 11 to 19 days longer under dense clumps of trees than under isolated trees in the Sierra (Figure 2), likely because of a combination of enhanced shading from solar radiation and deeper snow drifts. Patchy, shallow snow cover resulted in damped temperature oscillations, as observed in December under the single tree in Tuolumne Meadows and Tioga Pass (Figures 2b and 2c). Variation in the date of snow disappearance between sites (as much as 90 days) was larger than that within any particular site. Hodgedon Meadow lies in the intermittent snow zone, and snow cover appeared and disappeared multiple times through the winter.

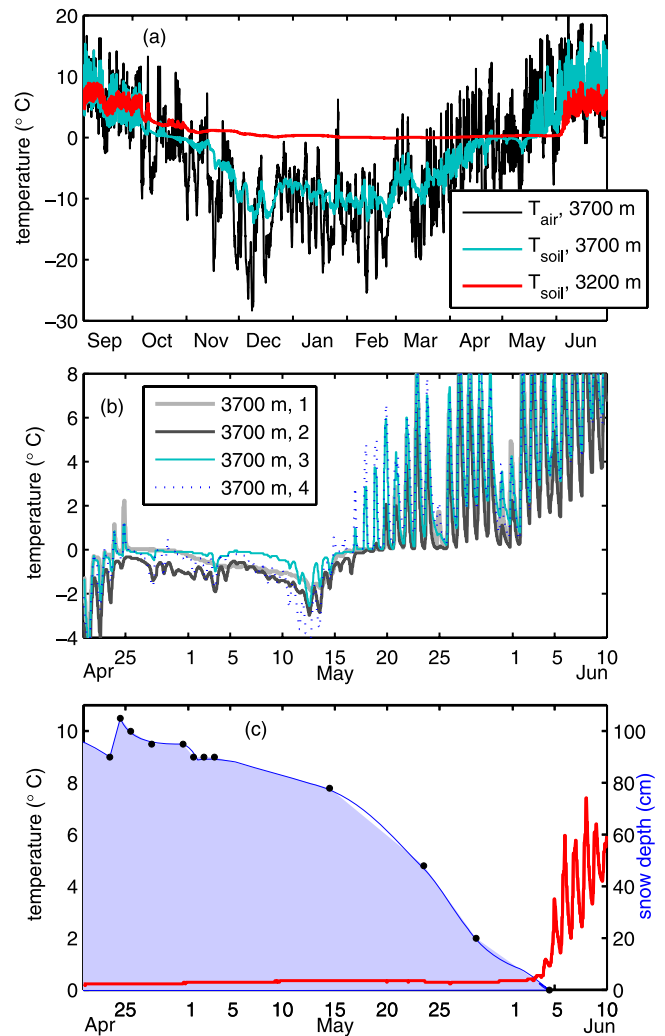


Figure 3. (a) Air temperature near D1 on Niwot Ridge (3700 m) and soil temperature both near D1 (3700 m) and about 2 km away below the tree line (3200 m) for September 2005 to June 2006. (b) Zoomed-in view of the snowmelt season, showing four soil temperature loggers at 3700 m near D1. No snow cover was present at these four sites on 20 April or 22 May 2006. Snow depths were 28, 10, 8, and 10 cm at sites 1, 2, 3, and 4, respectively, on 4 May 2006. (c) For the period shown in Figure 3b, soil temperature at the 3200 m site below the tree line (solid line, left axis), along with manual snow depth measurements (dots connected by line, right axis).

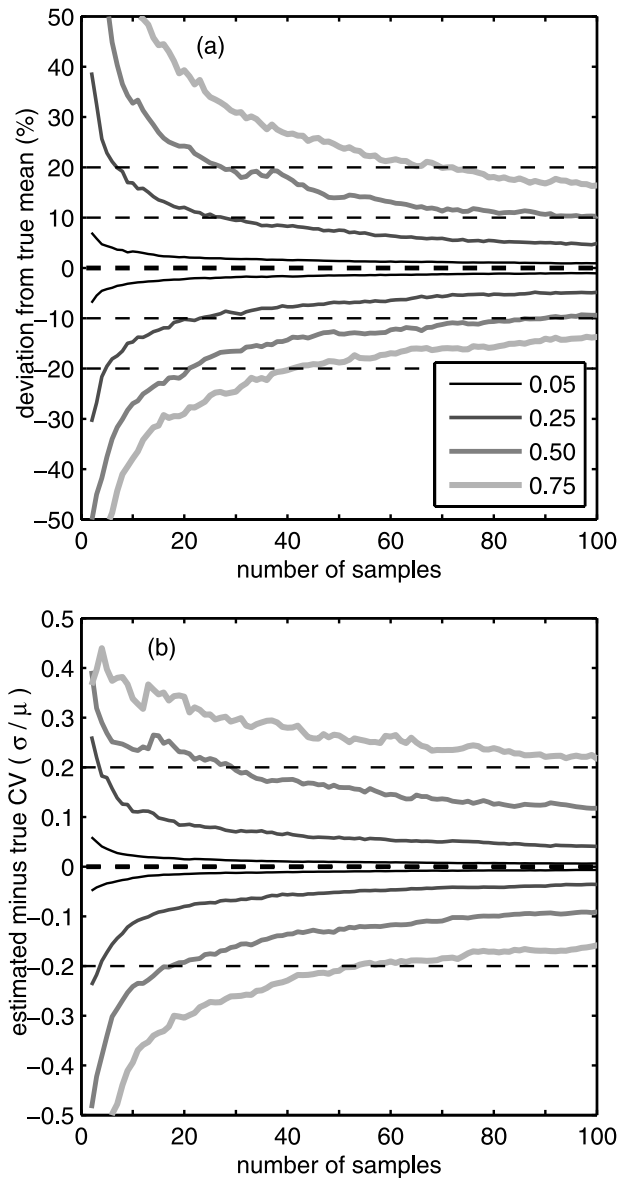


Figure 4. (a) The 95% confidence intervals of the percent difference between the estimated mean and the true mean for a lognormal distribution as a function of the sample size for coefficients of variation ranging from 0.05 to 0.75. (b) The 95% confidence intervals for the estimated minus the true coefficient of variation (CV) for the same distributions and legend shown in Figure 4a.

[12] On Niwot Ridge, Colorado, four pairs of air and near-surface soil temperature sensors were placed within 350 m of the D1 long-term climate station (elevation 3700 m, site description and map can be found in work by Litaor *et al.* [2008]). An additional temperature pair was located below tree line about 4 km away at an elevation of 3200 m. Air temperatures at the lower-elevation site were consistently about 3.5°C warmer than at the alpine sites, but the winter soil temperatures at the lower-elevation site were over 10°C warmer (Figure 3a). This occurred because at the sites above tree line on Niwot Ridge, snow depths were quite shallow (maximum measured depths ranged from 20 to 35 cm at the four sites), and wind blew the snow away,

exposing the soil surface, in between snowfall events. Here, frozen soils tracked air temperature but with damped diurnal oscillations when shallow snow was present (Figure 3a). During the spring (Figure 3b), manual measurements reported no snow cover at any of the four sites near D1 on 20 April 2006. On 4 May 2006, snow depths were 28, 10, 8, and 10 cm at sites 1, 2, 3, and 4, respectively, and no snow cover remained on 22 May 2006. The soil temperatures during this period indicate that new snow fell on about 25 April (damping of diurnal fluctuations) and melted between 16 and 20 May (resumption of diurnal fluctuations, Figure 3b). Likely because of its deeper snow cover, site 1 had the least diurnal variation during early May (Figure 3b). As in the arctic [Taras *et al.*, 2002], near-surface soil temperatures rose to 0°C when snow was actively melting (Figure 3b). At the site below tree line, which had snow depths of about 100 cm through most of the winter, soil temperatures were consistently 0°C when snow was present (Figure 3c). The date of snow disappearance varied by less than 3 days between the four alpine sites but by 15 to 18 days between the alpine and forested sites (Figures 3b and 3c). At all sites, once snow cover was gone, soil temperatures oscillated with diurnal cycles in temperature, providing a clear record of when snow was or was not present at a site (Figures 2 and 3).

5. Mapping Snow Disappearance Dates to SWE

[13] We used nearby CA DWR temperature and precipitation measurements to run a temperature index–based snowmelt model (Snow-17) [Anderson, 1973] at Tuolumne Meadows and Tioga Pass. We assumed that variations in snow disappearance date were due to variations in snow accumulation and adjusted the scaling of precipitation (the SCF parameter in Snow-17) for each point until the snowpack disappeared on the observed date. At Tuolumne, 5% more precipitation, resulting in 5 cm more SWE, was required at the dense stand of trees than at the single tree. At Tioga Pass, 25% more precipitation, resulting in 36 cm more SWE, was required at the dense stand of trees. The increase in variation at Tioga Pass likely occurred because the single tree was near a mountain pass and exposed to high winds, whereas in Tuolumne Meadows, the single tree was in a wind-sheltered opening.

6. Deployment Strategies: Which Sensors and How Many

[14] The resolution at which one should measure snowpack variability depends on the problem being addressed. Explicitly resolving the structure of the variability (e.g., its fractal scaling and precise mapping on the landscape) requires multiple transects of data points with meter-scale spacing from which to construct a variogram [Deems *et al.*, 2006]. However, for most water resources and meteorological applications, this level of detail is unnecessary. Subgrid-scale model parameterizations of snow depth, SWE, or SCA require knowledge of the mean and overall variance within a grid cell. Premelt distributions of SWE have often been approximated with a lognormal distribution [Essery and Pomeroy, 2004], and this has been used to parameterize subgrid cell rates of snow depletion [Luce *et al.*, 1999; Luce and Tarboton, 2004]. Pomeroy *et al.* [1998] report that the

coefficient of variation, CV, which is the standard deviation divided by the mean, varies from less than 0.05 in forests to more than 0.50 in open prairies. To test how many samples from a lognormal distribution are necessary to approach the true mean and variance, we used a random number generator to obtain 1000 samples from a lognormal distribution with a specified mean of 1 m and with CVs ranging from 0.05 to 0.75. We then estimated the mean and CV from sample subsets of lengths from 1 to 1000. After repeating this procedure 1000 times, we calculated 95% confidence intervals for the estimated mean and CV as a function of the number of samples taken and the CV of the distribution (Figure 4).

[15] The number of samples necessary depends on the accuracy required and the variability of the distribution. The greatest improvement in accuracy per additional sensor occurs with the first five sensors, with increasingly marginal improvements as additional sensors are added (Figure 4). These results are consistent with *Rice et al.'s* [2007] study with distributed snow depth sensors near a forest clearing, which demonstrated that four sensors reduced the uncertainty in a grid cell's actual average snow depth by over 40%, but the improvement with additional sensors was marginal. Thus, approximately 2–3 sensors could be randomly distributed within a 100-m forested grid cell (CV \approx 0.05) and 25–30 sensors within a grid cell in a wind-blown area above tree line (CV \approx 0.50) to estimate each grid cell's mean value within $\pm 20\%$ and its coefficient of variation within ± 0.2 (Figure 4). However, in areas of extreme variability, e.g., CV $>$ 0.50, many more sensors would be needed to characterize the local variability with the same accuracy.

[16] Most applications using snow depth and SWE must function at the watershed scale, which includes many hillslopes, each with likely different hillslope-scale variability, as well as a wide range of energy dynamics. For example, the snow cover duration varied far more between a site in the intermittent snow zone (Figure 2a) and sites at higher elevations in the Sierra Nevada (Figures 2b and 2c) than it varied within any pair of sites at a given elevation. Likewise, sites above and below tree line on Niwot Ridge exhibited much larger variability than was observed within the four alpine sites (Figure 3). Thus, for complex terrain, a cluster of sensors per elevation zone or vegetation type would be necessary to capture an approximation of both the local and larger-scale snow characteristics. As in work by *Molotch and Bales* [2006], physiographic variables, such as vegetation type and density, slope, etc., could be used to classify grid cells likely to have similar snow characteristics, and clusters of temperature sensors at one of each classification type could provide parameters for all similarly classified locations. Because each elevation zone would likely require a separate DTS logging system, this deployment strategy would be much better suited for self-recording temperature sensors.

[17] Table 1 describes the characteristics of both self-recording temperature loggers and DTS logging to elucidate which method would be preferable for various observational objectives. The self-recording sensors presented here are most practical for applications requiring fewer than 1,000 temperature measurements, with manpower available for deployment and retrieval, which do not require data in real

time. For example, these soil temperature sensors could be distributed in any configuration to test hypotheses ranging from the timing of snow cover in different physiographic zones, such as on different slopes or aspects, to spatial patterns of snow-vegetation feedbacks, such as relates to seedling growth and development. On the other hand, a precise calculation of the coefficient of variation in a wind-swept environment (CV $>$ 0.5) would require many more measurement locations and would be better suited for DTS logging.

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