Arctic snow distribution patterns at the watershed scale

Joel W. Homan and Douglas L. Kane

ABSTRACT

Watershed-scale hydrologic models require good estimates of spatially distributed snow water equivalent (SWE) at winter's end. Snow on the ground in arctic environments is susceptible to significant wind redistribution, which results in heterogeneous snowpacks. The scarcity and quality of data collected by snow gauges provides a poor indicator of actual snowpack distribution. Snow distribution patterns are similar from year to year because they are largely controlled by the interaction of topography, vegetation, and consistent weather patterns. Consequently, shallow and deep areas of snow tend to be spatially predetermined, resulting in depth (or SWE) differences that may vary as a whole, but not relative to each other. Our aim was to identify snowpack distribution patterns and establish their stability in time and space at a watershed scale. Snow patterns were established by: (1) using numerous field surveys from end-of-winter field campaigns; and (2) differentiating snowpacks that characterize small-scale anomalies (local scale) from snowpacks that represent a large-scale area (regional scale). We concluded that basic snow survey site descriptions could be used to separate survey locations into regional and local-scale representative sites. Removing local-scale influences provides a more accurate representation of the regional snowpack, which will aid in forecasting snowmelt runoff events.

Key words | Alaska, arctic snow distribution, heterogeneous snowpacks, snow water equivalent, watershed scale

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INTRODUCTION

Snow hydrology is an important component of the arctic hydrologic cycle. In Alaska, the north-flowing rivers of the Arctic drain three physiographic areas: Mountains, Foothills, and Coastal plain. The total water content of the snowpack at the end of winter within the Arctic comprises 30 to 40% of the annual precipitation (Kane *et al.* 1991, 2008b), and on average, about two-thirds of the snow water equivalent (SWE) leaves catchments as runoff (Kane *et al.* 2000, 2004, 2008a). On the other hand, the average runoff ratio for rainfall events is roughly one-third for most summer precipitation events (Kane *et al.* 2012). The higher runoff ratio in spring is mostly due to generally frozen subsurface conditions of the active layer. The low runoff ratio of the summer months is partially due to roughly 140 mm of evapotranspiration over the basin (Kane *et al.* 2004) and surface storage availability on the low-gradient Alaskan Coastal Plain, where wetlands and lakes cover 82.9% of the landscape (Hall *et al.* 1994). There is also a disparity in the watershed areas that can contribute to runoff and subsequent runoff ratio. During snowmelt, the entire basin potentially contributes melt water, while summer rainfall events generally only occur over a portion of the watershed (Kane *et al.* 2008b).

Heterogeneous snowpacks resulting from snow redistribution is also a major factor in increasing the snowmelt runoff. Much of the redistributed snow accumulates in valley bottoms, in hillside depressions such as shrubby water tracks, and on leeward sides of ridges. Snow drift formation in hillside depressions and valley bottoms generally results in proportionally higher water content close to or within drainage channels, which potentially increases runoff. The combination of above-average water content

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close to the drainage channels, reduced time of transport of hillslope meltwater due to water tracks, and a completely frozen active layer that limits subsurface melt water storage produces a generally higher runoff ratio during snowmelt than is observed for rainfall runoff events (McNamara *et al.* 1998). Some extreme summer storms, however, have produced flows greater than the largest measured snowmelt flood, but only in small high-gradient headwater basins. For large arctic river basins here and elsewhere, the spring snowmelt floods dominate and can be expected every year (Kane *et al.* 2008b).

In the Alaska Arctic, where snow accumulation may last for nine months and then ablate in a relatively short time, typically 10 to 14 days just before the solstice, the end-of-winter SWE plays a significant hydrologic role in watersheds (Kane & Hinzman 1988). The task of accurately quantifying solid precipitation in the Arctic is made difficult because it is a remote, sparsely inhabited, and severely cold environment. Snowfall itself is a stochastic process, and variability is inevitable. Other than the inherent variability in snowfall, an additional problem is that often the quality of precipitation measurements from meteorological stations is poor. For instance, in the Alaska Arctic, snowfall precipitation has been shown to be underestimated by a factor of two or three when windy conditions prevail (Benson 1982). Redistribution of snow by wind can create complex snow distributions resulting in heterogeneous or even patchy snowpacks (Elder *et al.* 1991; Seyfried & Wilcox 1995; Prasad *et al.* 2001; Winstral *et al.* 2002; Anderton *et al.* 2004; DeBeer & Pomeroy 2009).

The redistribution of snow may be complex, but generally forms consistent patterns. Since snow crystals behave similarly to other sediments, it tends to accumulate in areas where flow diverges or decelerates, while it erodes in areas of convergent or accelerated flow (Elder *et al.* 1991). Because of this, snow distribution patterns are similar from one year to the next because they are largely controlled by the interaction of topography, vegetation and consistent synoptic weather patterns (Sturm & Wagner 2010). The number of wind events, wind magnitude and direction, vegetation density and canopy height, and topography (aspect, slope and elevation) are all factors that are important to the end-of-winter snowpack distribution (Elder *et al.* 1991; Pomeroy & Gray 1995; Konig & Sturm 1998; Winstral *et al.* 2002; Sturm & Wagner 2010). Difficulties in measuring falling solid precipitation, as well as quantifying snow redistribution and winter-long sublimation, make ground-based snow surveys at winter's end the most reliable and economical approach for quantifying the SWE that will contribute to runoff.

The primary goal of this study is to provide better definition of the distribution of solid precipitation data for input into hydrologic models. Specifically, the intention is to improve the understanding of watershed-scale spatial variability of solid precipitation at winter's end in the central region of the Alaska Arctic including the Dalton Highway corridor (the only major transportation route). Snow survey data used for this investigation were collected from 2000 to 2013 by faculty, staff, and students at the Water and Environmental Research Center (WERC) at the University of Alaska Fairbanks (UAF). From the data collection, snowpack distribution patterns can be examined over the entire region.

Warm season precipitation patterns within a section of the study area (Kuparuk River watershed) were evaluated in 1993 and 1994 by Kane *et al.* (2000). Their findings indicate a direct correlation between rainfall and elevation, with increasing precipitation from lower to higher elevations. While summer precipitation patterns were evident with elevation, the distribution of winter SWE was less conclusive. Most of these uncertainties arise from a sparse observational network and the short period of observation at the time of the study by Kane *et al.* (2000). This paper presents new findings using a long-term, widely distributed (2000–2013) snow dataset.

STUDY DOMAIN

The study domain covers a 200 by 240 km region of Alaska's Central Arctic Slope that is bound by the Brooks Range on the south and the Arctic Ocean on the north, and includes the Chandler, Anaktuvuk, Itkillik, Kuparuk, Putuligayuk (Put), Sagavanirktok (Sag), Kadleroshilik (Kad), and Shaviovik River basins (Figure 1). All of the watersheds drain north and eventually empty into the Arctic Ocean or another stream that eventually discharges to the ocean. The Putuligayuk lies entirely within the

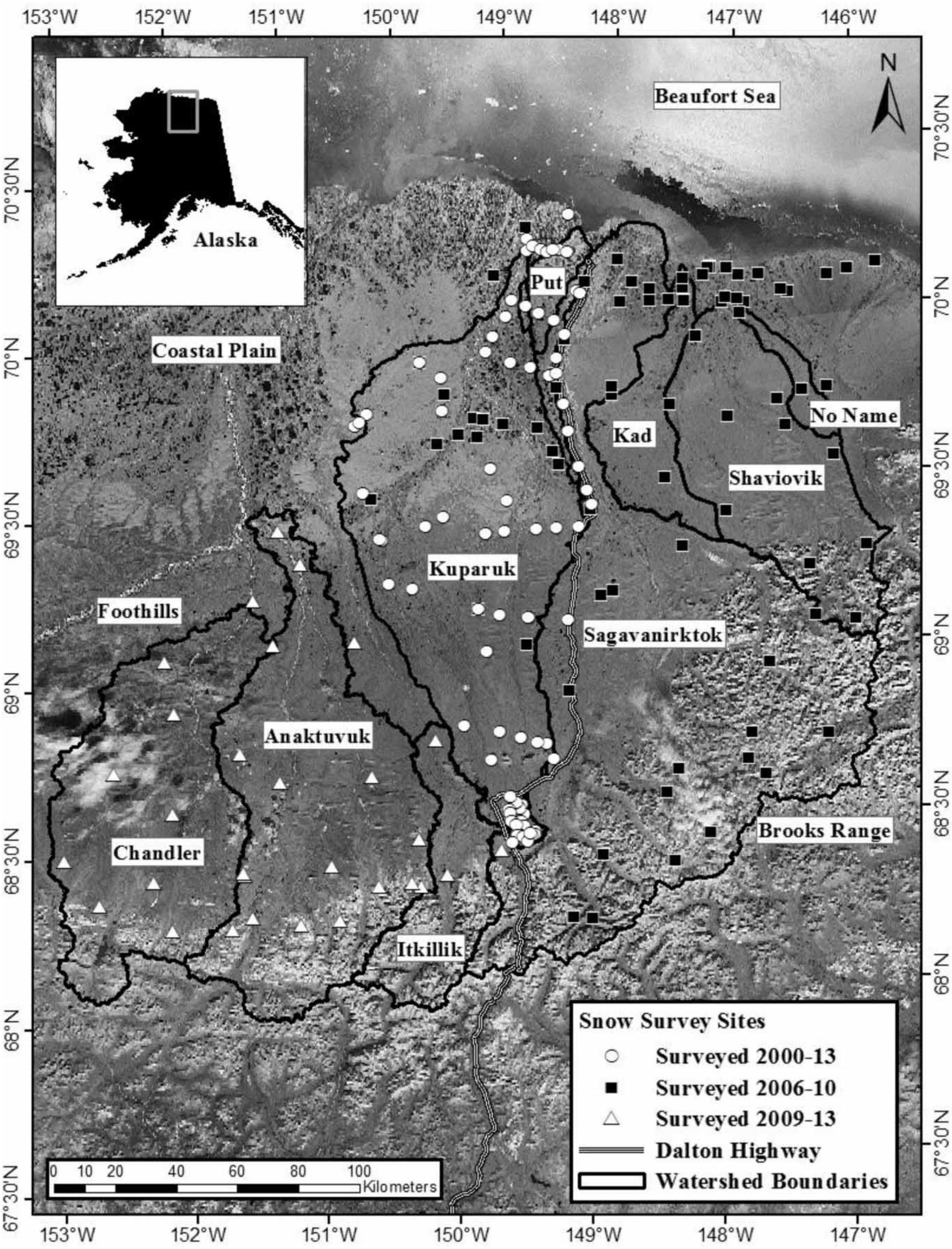

Figure 1 | Site map showing snow survey locations within several Central Alaska North Slope watersheds.

Coastal plain region; the Kuparuk and Kadleroshilik Rivers emanate from the Foothills and cross the Coastal plain; the Sagavanirktok, Shaviovik, Kavik, Itkillik, Anaktuvuk, and Chandler Rivers originate in the Brooks Range and cross both the Foothills and Coastal plain. The southern and northern boundaries of the domain are at between 68°N and 70°N latitude, while the western and eastern boundaries are between 153°W and 146°W longitude. Elevation within the study area ranges from sea level to 2675 m. The topography is characterized by a flat northern portion (generally referred to as 'Coastal Plain') and by gently rolling hills and valleys ('Foothills') and mountain ridges ('Mountains') of the Brooks Range to the south.

The entire study area is underlain by continuous permafrost (250 to 300 m in the Brooks Range and up to 600 m along the coast; Osterkamp 1984) and, on average, is snow-covered for 8 to 9 months of the year. The region is mostly treeless with some patches of trees in the riparian areas in the Foothills. Vegetation consists of alpine plant communities in the mountainous region, tussock tundra in the Foothills, and sedge tundra on the Coastal Plain (Walker *et al.* 1989; CAVM Team 2003). Willow and birch shrubs are common in riparian areas, and shrub height is variable, from approximately 0.3 to over 1 m. In response to climate warming in the Arctic, there is an increase in the abundance and extent of shrubs in tundra areas (Sturm *et al.* 2001, 2005; Tape *et al.* 2006).

SURVEY LOCATIONS

From 2000 to 2013, over 1000 snow surveys were conducted at roughly 200 locations. The snow survey dataset is a collection of results from numerous research projects, so the exact number of sites surveyed and their locations changed yearly. The distribution of snow survey sites is shown on a map of Alaska's Central North Slope (Figure 1). The symbol classifications best describe the timing and duration of surveys, but they do not mean that the sites were visited every year within each classification. Ideally, sites would have been surveyed every year throughout the duration of the projects. Weather conditions, however, play a large role in the feasibility of reaching most of the remote survey sites, which are accessible only by snow machine and/or helicopter. High wind, fog, whiteout, and flat light conditions prevented some of the sites from being surveyed every year, although the goal was to monitor them if safely possible.

The snow survey sites were chosen to represent a wide range of snowpack conditions. Initially, from 2000 to 2005, the National Science Foundation (NSF) funded snow surveys that were primarily within the Kuparuk and Putuligayuk (Put) River watersheds (Figure 1). In 2006, under an Alaska Department of Natural Resources (ADNR) project, snow surveys were extended eastward to the Sagavanirktok, Kadleroshilik, and Shaviovik River watersheds (Figure 1). Also in 2006, additional sites were added in the Kuparuk River watershed on a project funded by the Alaska Department of Transportation and Public Facilities (ADOT&PF) (Figure 1). Finally in 2009, data collection progressed further westward to the Anaktuvuk, Itkillik, and Chandler River watersheds under the ADOT&PF project (Kane *et al.* 2012) (Figure 1).

Surveying along a uniform grid would have been inadequate at a watershed scale because the Arctic snowpack is very heterogeneous (Kane *et al.* 1991; Homan *et al.* 2010; Sturm & Wagner 2010), with relatively shallow snow on hilltops, along ridges, and on steep slopes, while deeper snow accumulates in valley bottoms and water tracks, and on leeward slopes. Elevation, terrain, vegetation cover, and spatial distribution were all considered during the survey site-selection process. Snowmelt studies have shown that areas with deeper snow take several days or weeks longer to completely melt compared with areas of shallower snow (Hinzman *et al.* 1991). In the end, the snow survey sites were positioned to represent both 'regional' and 'local' snow conditions in order to capture a greater spatial variability of the snowpack. The regional-scale sites are more uniform and characteristic of larger-scale snow conditions (1 to 10 km^2), while local-scale sites are smaller, tend to be linear and represent more limited features such as wind-scoured ridges (10 to 1000 m) or snowdrift (1 to 10 m) deposits within depressions such as streams and water tracks.

SNOW SURVEY METHODS

The snow surveys include snow density sampling and snow depth measurements collected over an area of 25 m by 25 m;

this technique is often referred to as double sampling (Rovansek *et al.* 1993). The snow depth of the snowpack in Alaska is more variable than the density (Benson & Sturm 1993; Sturm *et al.* 2010). Usually, double sampling yields an areal SWE estimate with a lower variance than is possible using collected snow cores only. In addition, considerably more snow depth measurements can be made in a unit of time compared with SWE measurements. Rovansek *et al.* (1993) showed that double sampling provides improved SWE estimates; they recommended sampling 12 to 15 snow depths for each snow core. This optimal ratio of snow depths to water equivalent, however, appears to vary greatly (from 1 to 23), depending on site, weather, and snow conditions. The UAF-WERC uses an optimal ratio of 10; that is, 50 depths accompany five snow cores at each survey site.

Snow cores are sampled using a fiberglass tube ('Adirondack') with an inside area of 35.7 cm^2 (diameter = 6.7 cm) and length of 152.4 cm (5 ft), equipped with metal teeth on the lower end to cut through dense layers of snow. The advantage of the Adirondack for shallow snowpack is that its diameter is larger than many other types of snow tubes (like the Mt. Rose); thus, it provides a larger sample of the shallow Arctic snowpack. To obtain a complete snow core, the Adirondack tube is pushed vertically through the snow while turning, until soil is encountered. At this point, snow depth is recorded. The tube is then driven further into the organic layer and tipped sideways, retaining a vegetation plug. This extra step is important, because the base of the Arctic snowpack generally consists of poorly consolidated depth hoar and the organic plug acts as a stopper, ensuring that the complete snow column was sampled. The tube is then removed from the snowpack and the vegetation plug is discarded. The snow itself is either collected for weighing later in the laboratory or weighed immediately in the field.

The WERC uses constant 50 m lengths for the snow depth course, with a 1 m sampling interval along an L-shaped transect (Kane *et al.* 2012). Twenty-five depth measurements are made on each leg of the L; this strategy is used to account for the presence of snowdrifts in the area of measurement. The directions of measurement are chosen randomly. Snow depth measurements are made using a T-shaped graduated rod (T-probe). The probe is simply pushed through the snow to the snow–ground interface. The SWE is defined as follows:

$$\mathrm{SWE} = SD*(\rho_s/\rho_w) \quad (1)$$

where SD is an average of 50 snow depths, ρ_s is average snow density from the five snow core samples, and ρ_w is water density (Kane *et al.* 2012; Stuefer *et al.* 2013). Snow depths are reported in centimeters (cm), while SWE is reported in millimeters (mm).

RESULTS

Winter in the Alaskan Arctic starts with snowfall and ends with snowmelt and subsequent runoff. In the Arctic, snow can fall any day of the year, but snow accumulation typically begins in September or early October and continues throughout the entire winter, with no significant midwinter melt. It is hypothesized, however, that both midwinter melt and rainfall are more likely as the climate warms. Snow accumulation occurs during a few large events, many small events, or somewhere in between, but more commonly from a variety of event sizes. It is incorrect to assume that snowfall at a given location is equivalent to snow accumulation recorded during field measurements or by a precipitation gauge, or that a location's measured accumulation can be easily extrapolated to another location. Snow on the ground in treeless arctic environments is susceptible to significant wind redistribution, which is normally accompanied by in-transit sublimation. As a result of both transport and sublimation, the end-of-winter snowpack is very heterogeneous, and depending on the survey location, the end-of-winter snowpack can be as thin as a dusting or can exceed a meter in depth.

Snowpack distribution

Though the snowpack itself may be heterogeneous, the redistribution of snow has been found to have some consistencies: snow is depleted from hilltops, ridges, and windward slopes, while it accumulates in valley bottoms, on leeward slopes, in any depressions such as water tracks and between hummocks, and within or around vegetation.

To identify snowpack distribution patterns, SWE measurements from the long-term (2000–2013) snow dataset were used (Table 1). The 14-year dataset has a mean density of 246 kg/m^3 (Figure 2(a), Table 2) and exhibits a nearly symmetric distribution, while depth (mean = 42 cm) (Figure 2(b), Table 2) and SWE (mean = 103 mm) (Figure 2(c), Table 2) have asymmetric distributions. The similarity between the SWE and depth frequency curves is indicative of the fact that SWE is more closely linked to depth than it is to density. This relationship is further verified by plotting SWE vs. depth (Figure 3(a)), which has regression lines with R^2 values in the 0.80 s. The correlation values are high, and the slopes of the regressions are equal to the mean densities, which they should be, given the equation of SWE (Equation (1)). In comparison, a plot of density versus SWE (Figure 3(b)) shows considerably more scatter and little to no correlation.

Due to the physical processes that lead to the densification of snow, it is known that there is a density to depth relationship, where greater snow depths have on average higher densities (Kojima 1966). This is not the case with the current dataset, which illustrates almost no change in

Table 1 | Yearly (2000–2013) end-of-winter snowpack characteristics for the entire study domain; including sample size, mean, standard deviation (Stdev), range, minimum and maximum

	Depth cm	Density kg/m^3	SWE mm	Depth cm	Density kg/m^3	SWE mm	Depth cm	Density kg/m^3	SWE mm	Depth cm	Density kg/m^3	SWE mm
Year		2000			2001			2002			2003	
Sample size		65			85			85			86	
Mean	47	236	111	44	237	103	43	238	100	43	264	113
Stdev	13	26	35	16	36	38	15	38	36	14	41	40
Range	70	167	181	79	198	205	82	208	160	78	260	210
Min	14	168	30	14	183	29	12	151	39	14	179	33
Max	84	335	210	93	380	235	94	359	199	91	439	243
Year		2004			2005			2006			2007	
Sample size		56			81			118			150	
Mean	42	259	107	38	251	96	38	231	88	39	238	92
Stdev	13	39	36	17	37	44	13	49	34	17	45	43
Range	69	190	191	75	202	218	75	242	186	92	237	311
Min	15	182	39	8	137	21	11	137	21	1	141	3
Max	84	371	230	82	339	239	87	380	208	93	378	314
Year		2008			2009			2010			2011	
Sample size		106			143			104			77	
Mean	34	237	83	50	258	126	39	260	101	47	248	119
Stdev	16	49	46	22	48	61	15	40	41	22	39	65
Range	101	269	328	122	255	399	73	187	221	118	210	356
Min	3	138	5	8	157	14	5	155	15	6	153	23
Max	103	408	334	129	412	413	78	342	236	124	363	380
Year					2012			2013				
Sample size					73			76				
Mean				40	258	104	47	237	112			
Stdev				15	39	42	18	36	47			
Range				72	203	221	84	177	240			
Min				3	152	6	4	159	12			
Max				75	356	228	87	336	252			

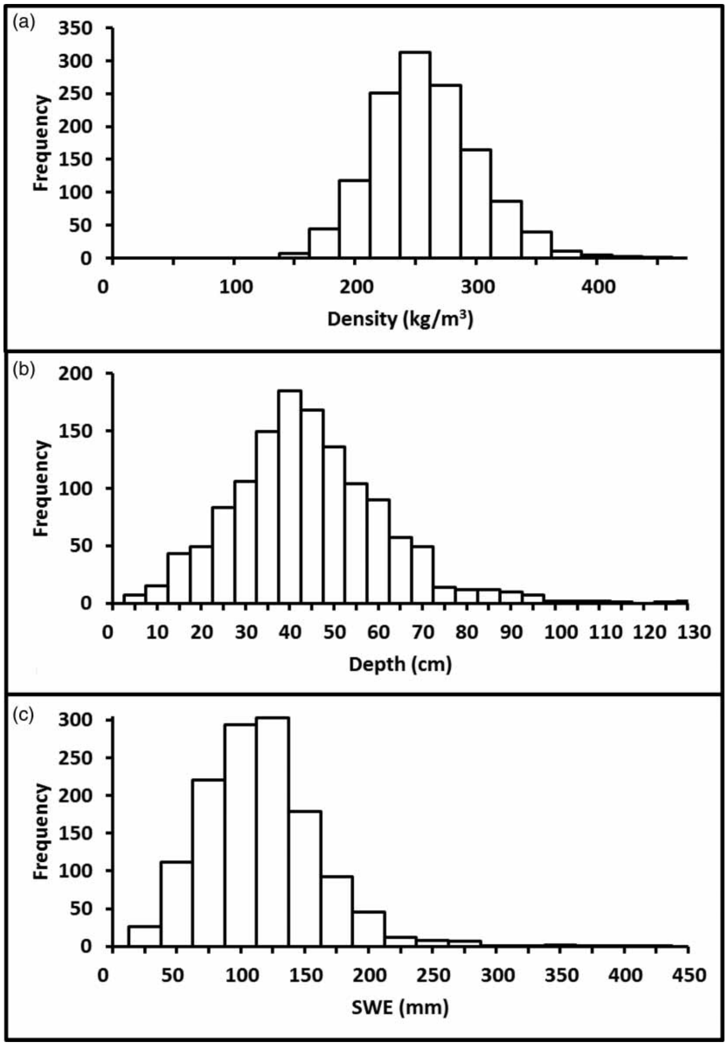

Figure 2 | Frequency distributions for the complete record of snow densities, depths, and SWEs. Densities have a nearly symmetric distribution, while snow depth and SWE have asymmetric distributions.

density with depth (Figure 3(c)). The discrepancy is most likely a result of relatively shallow arctic snowpacks that do not reach sufficient compaction depths.

Within the dataset are distinctly different patterns for two different regions. One pattern exists at lower elevations (Lowlands <225 m) and another at higher elevations (Uplands >225 m) (Figures 3–5, Table 2). The Uplands snow has on average a lower density (Figures 3(b) and 3(c), Figure 4(a)) which is attributed to inland increases in thickness of the depth-hoar layer where greater temperature extremes (in particular, lower minimum temperatures) permit larger gradients to develop within the snow-pack (Hall *et al.* 1986). The density for depth hoar layers varies from 150 to 250 kg m^{-3}, while the density for wind packed layers (wind slab) fluctuates from 400 to 500 kg m^{-3} (Benson & Sturm 1993). Overall, the Lowlands and Uplands snow densities were found to have poorly fit regressions with elevation (low R^2 values), which illustrates high variability (Figure 5). It was statistically proven that snow density-to-elevation relationships did not exist with significant probability (p-values less than 0.01, 99% chance the statistical relationships are not 'real'). The separation of this region into these categories is, however, supported by Sturm & Stuefer (2013). Their findings showed that both wind speed and the number of wind events are greater along the coast and decrease from the Lowlands towards the Uplands,

Table 2 | Record average end-of-winter (2000–2013) snowpack characteristics for the entire study domain, Lowlands and Uplands; including sample size, mean, standard deviation, range, maximum and minimum

	Study domain			Lowlands (<225 m)			Uplands (>225 m)		
	Depth cm	Density kg/m^3	SWE mm	Depth cm	Density kg/m^3	SWE mm	Depth cm	Density kg/m^3	SWE mm
Sample size		1305			631			674	
Mean	42	246	103	41	256	104	43	237	102
Standard deviation	17	43	47	15	45	44	19	39	50
Range	128	302	410	127	275	364	128	301	410
Min	1	137	3	3	137	5	1	138	3
Max	129	439	413	129	412	369	129	439	413

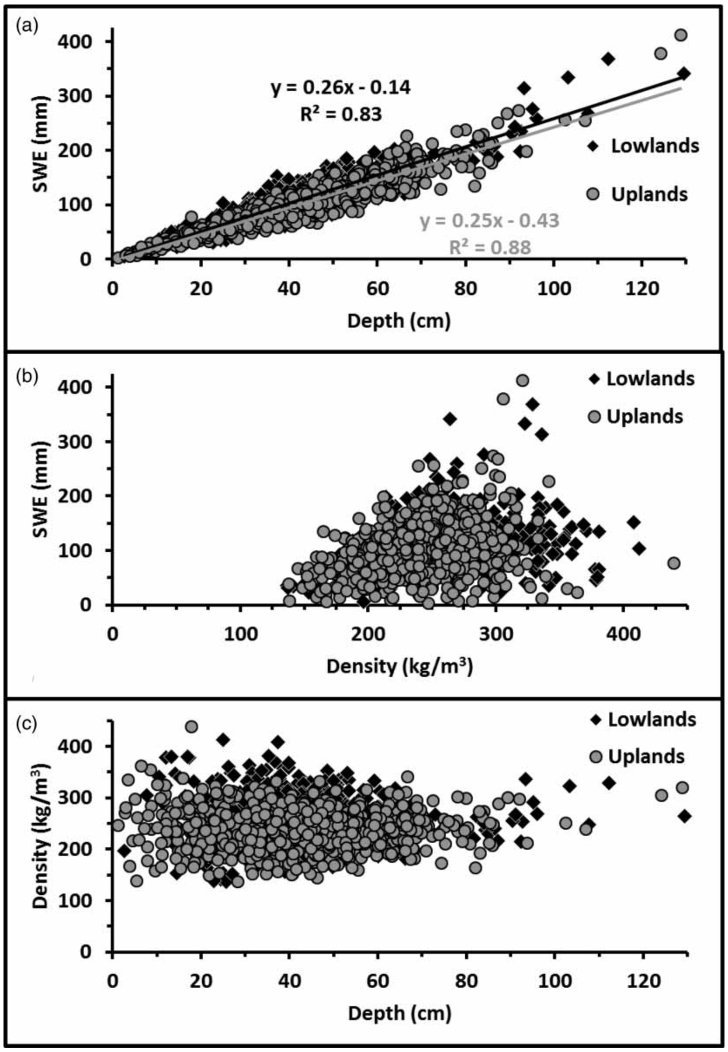

Figure 3 | Scatter plots to assess similarity between SWE and snow depths compared with density.

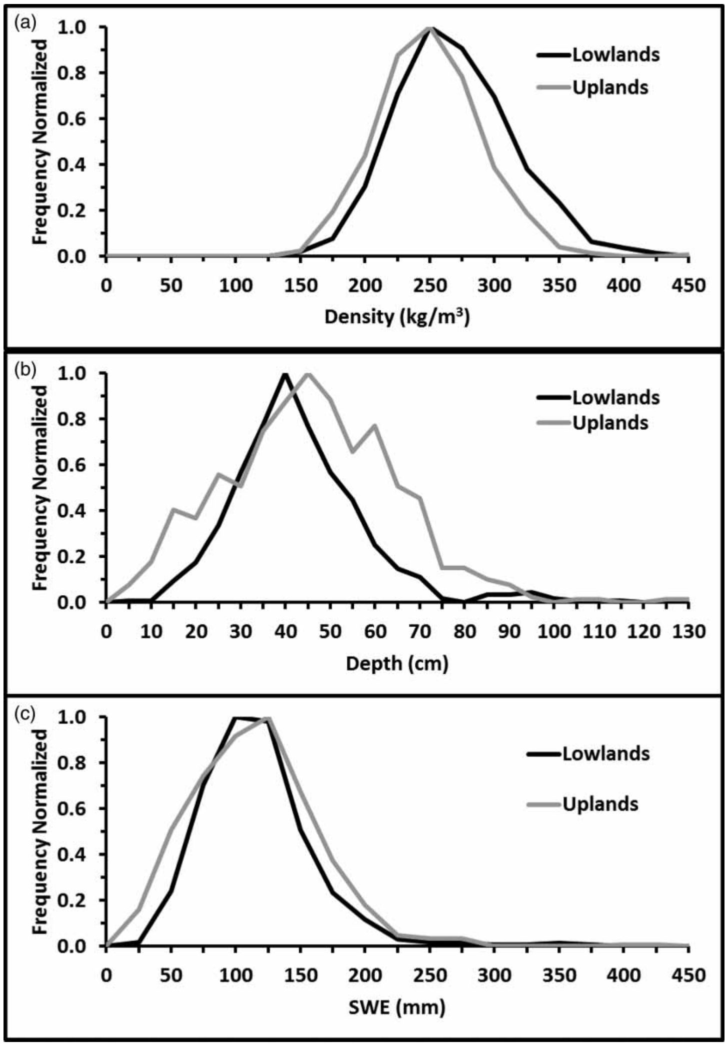

Figure 4 | Probability distribution plots to demonstrate the differences in snow densities, depths, and SWEs between the Lowland and Upland regions.

which could also facilitate higher Lowland densities by developing a greater extent of wind slab.

Other than higher densities, the Lowlands also have shallower and less variable snow depths and SWEs (Figures 4(b) and 4(c), Table 2). Initially, both snow depth and SWE slightly increase with increasing elevations, indicating a weak orographic effect near the ocean (Figures 5(b) and 5(c)). However, at elevations greater than 225 m, an orographic effect is no longer apparent, which is believed to be a result of decreasing moisture content with distance from the ocean (Liston & Sturm 2002; Kane *et al.* 2012). The variability of snow depths and SWEs are high for both the Lowlands and Uplands and as a whole, elevation relationships were statistically proven to not exist with significant probability (p-values <0.01).

Overall, at a watershed scale, the survey data demonstrate a slight decrease in SWE with increasing elevation (Figure 6). More specifically, there is a decrease of only 9.4 mm of SWE for every 1000 m of elevation gain. This is a significant difference compared with summer precipitation, which increases more than 200 mm for the same 1000 m elevation gain (Kane *et al.* 2000). Essentially, the present dataset illustrates that the average SWE at winter's end is relatively independent of elevation on Alaska's Central North Slope, with roughly an overall average of 10 cm of water from the Coastal Plain to the Mountains.

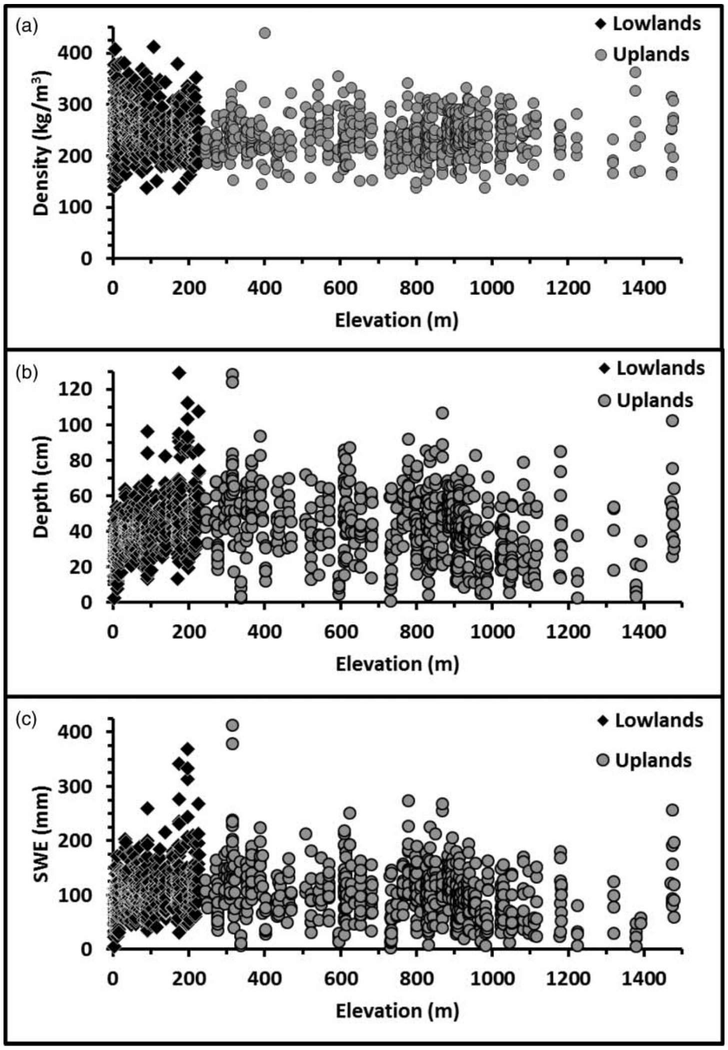

Figure 5 | Distribution curves to illustrate the correlation between snow densities, depths, and SWEs to elevation.

All of the measured SWE values represent a certain percentage of Alaska's Central North Slope. Some SWE values might characterize small-scale anomalies (local scale), while others might signify a large-scale area (regional scale). Since our interest was in identifying snowpack distribution patterns at a watershed scale, snow survey sites that represent local-scale snowpacks needed to be identified and removed. After analyzing the long-term dataset, it was concluded that basic snow survey site descriptions could be used to separate the survey locations into regional and local-scale representative sites. The decisions were made using on-site terrain, vegetation and snow pattern characteristics (i.e. windswept ridge, valley bottom depression, thick willows ~1 m tall, broad open and flat). Descriptions strongly analogous with snow drifting or scouring were classified as local-scale, while the remaining snow survey sites were left to represent the regional snowpack. Local-scale classifications were further divided into high and low water content subdivisions (i.e. wind-swept = Low SWE, drifted landscapes = High SWE). Fifty-five additional survey sites and surveys were not included as a result of only being surveyed for 1 year and lacking detailed site descriptions.

Of the classified snow survey sites, 25% represented local-scale snowpacks (Figure 6, Table 3). Most of the snow survey sites consistently represented their

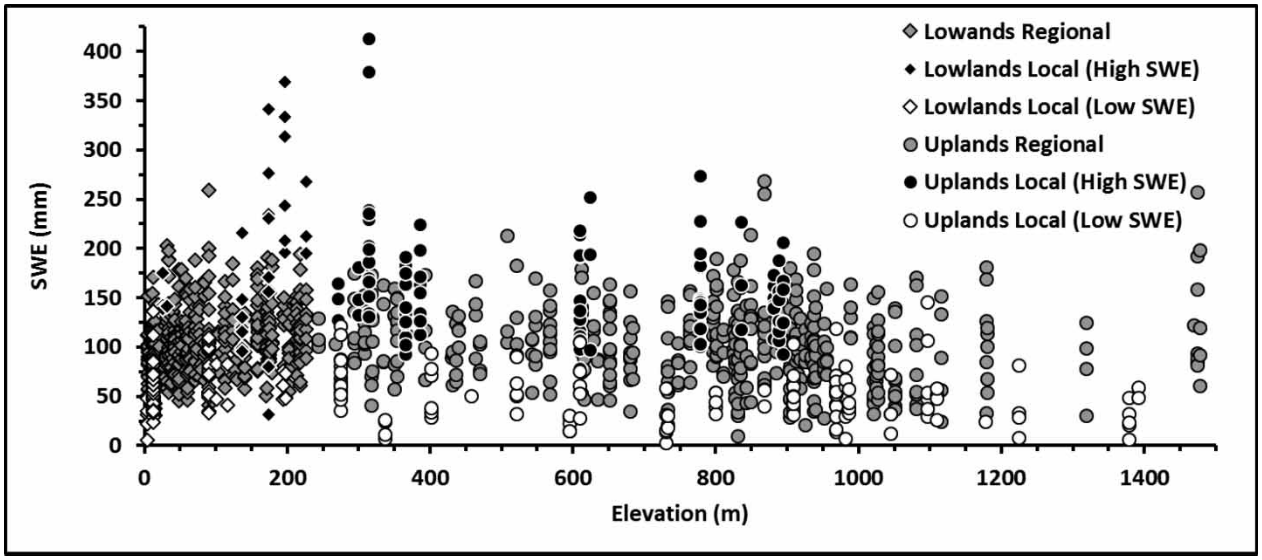

Figure 6 | Classified (high and low SWE local-scale and regional-scale) SWE values plotted against snow survey site elevation. Differentiation between Lowlands and Uplands is also indicated by symbol changes at the 225 m elevation boundary. Record averaged SWE from the Coastal Plain to the Brooks Range is 103 mm, with only a 9.4 mm decrease in SWE over a 1000 m elevation change. With the removal of local-scale outliers, the change in SWE was reduced to a 2.7 mm decrease for the same 1000 m elevation change.

Table 3 | Record snow survey information for the entire study domain, regional-scale and local-scale (high and low water contents subdivisions) representative areas; including number of survey sites and surveys, percentage of survey in each subdivision along with SWE averages and ranges

			Local-scale	
	Total/ average	**Regional-scale**	**high SWE**	**Low SWE**
# Survey sites	186	136	18	32
# Surveys	1250	940	124	186
% of Total surveys		75%	10%	15%
Average SWE (mm)	103	105	168	53
Range (mm)	410	259	381	143

pre-described classification (time after time having high or low SWE), but some locations occasionally have water contents outside their domain. The SWE measurements from local-scale features were removed to provide a more accurate representation of the regional snowpack. The removal of local-scale outliers had very little effect on the overall average SWE (changed from 103 to 105 mm) and trend line slopes (not shown in Figure 6). The slight change in slopes is a result of a greater percentage of drifting (high SWE) outliers in the Lowlands and wind-swept (low SWE) outliers in the Uplands. This tendency, which was expected, is a by-product of higher elevations having more areas that promote snow removal, while lower elevations have more depressions that enhance snow accumulation. For example, the mountains in the Upland region normally have more hilltops, ridges, steep slopes, and less vegetation, all of which characteristically are sites of snow erosion, while the Lowlands have more valley bottoms, surface water bodies, and vegetation for the snow to accumulate on and around. By using all of the SWE measurements, the snowpack distribution patterns can be obscured by local-scale influences. The removal of survey sites that represent local-scale anomalies provides a more accurate representation of the regional snowpack.

Yearly trend lines

Yearly trend lines were plotted to evaluate the stability of the snowpack in time and space (Figure 7). The lower elevations illustrate a tighter grouping of average SWEs, while an increase in elevation results in greater yearly variability. The greater variability at higher elevations results from increased complexity in slope and topography,

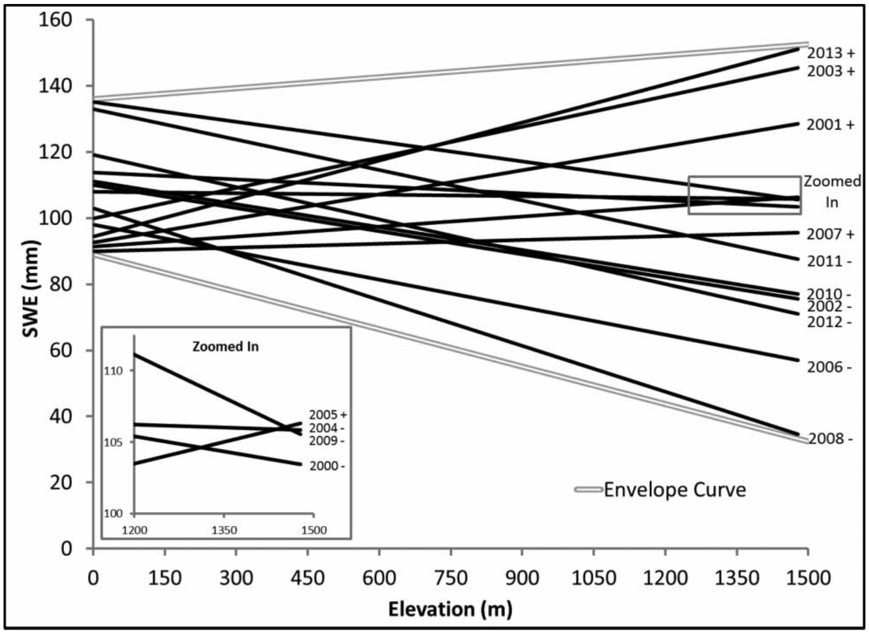

Figure 7 | Yearly trend lines using annual datasets for the 14-year record illustrate the temporal and spatial variability of the snowpack. Yearly slope directions (positive (+) or negative (−)) are specified after year of trend line labels. Nine years have negative trends with elevation, while 5 years have positive.

while lower elevations are generally more uniform. Nine of the 14 surveyed years had negative sloping trend lines, while 5 years had positive sloping trends.

DISCUSSION

Numerous experimental studies have been done on the distribution of snow (Martinec & Rango 1981; Elder *et al.* 1991; Benson & Sturm 1993; Sturm *et al.* 1995; Konig & Sturm 1998; Anderton *et al.* 2004; Stuefer *et al.* 2014). Most snowpack spatial distribution studies in mountainous catchments, such as Anderton *et al.* (2004) and Elder *et al.* (1991), suggest that topographic controls are the most important influence on snow distribution. This study also suggests that topography plays an important role in the distribution of snow within Alaska's Central Arctic. Within the study area, findings indicated that spatial variability greatly increases with increasing elevation. What was not found, however, was a linear increase of winter precipitation with increasing elevation, which is found in many regions (Golding *et al.* 1968; Singh & Kumar 1997; Hanson 2001).On the contrary, this study found a slight overall decrease in SWE from the Coastal Plain to the Mountains. Initially, an increase in winter precipitation occurs with increases in elevation, but that increase in precipitation is interrupted at 225 m. Above this level the amount of winter precipitation decreases with increasing elevation, which is analogous to increasing distance from the ocean. This finding correlates well with previous work, which has shown moisture content to decrease with distance from the ocean (Liston & Sturm 2002; Kane *et al.* 2012). Together these results suggest that in this region, winter precipitation is more dependent on available moisture rather than topographic controls and is under investigation for a subsequent paper.

In an attempt to identify snowpack distribution patterns and establish their stability in time and space at a watershed scale in Alaska's Central Arctic, small-scale anomalies were differentiated from regional-scale snowpack characteristics. Many of the snow-surveyed sites constantly represent either regional or local-scale snowpacks. The point-source sampling strategy was improved by sampling only locations that consistently provided SWEs that characterize the regional average snowpack, thus more representative data were collected.

CONCLUSIONS

The presently collected long-term snow dataset provides a rare opportunity to explore spatial- and temporal-scale variations over a large-scale area in the Alaskan Arctic. Using this dataset allowed the identification of a snowpack distribution pattern at a regional scale. More specifically, it was determined that SWE measurements are affected by the location in which the surveys were conducted. Some locations consistently produce either extreme high or low SWE values and represent local-scale snowpack irregularities. Through survey site descriptions and classifications, these local-scale high and low SWE representative sites were identified and removed. The remaining snow survey sites were left to represent what was classified as the regional snowpack. The removal of local-scale SWE measurements had little effect on the record average SWE, which, as concluded was relatively independent of elevation on Alaska's Central North Slope, with roughly an average of 10 cm from the Coastal Plain to the Mountains. At the smaller scale, however, slight elevation-based snowpack patterns were identified above and below 225 m elevation. At the watershed scale, the overall lack of variability in SWE with change in elevation was unexpected and inconsistent with summer precipitation patterns within this region and snow accumulation in most other regions. It is the authors' opinion that the discrepancy comes from this region being extremely moisture deprived during the frozen winter months and, therefore, that snowfall is dependent on moisture availability rather than topography.

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