Snow retention for increased water supply of shallow arctic lakes

Running title: How to gain a lot of water with a little snow?

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Abstract

Throughout the year, water availability in the Arctic is marginal because of the cold climate, presence of permafrost, and low precipitation. Recent resource development and improvements in living conditions of the expanding population have placed additional stresses on the annual local water supply of communities and industry located in this region. This study implemented a snow control practice to enhance snowdrift growth as a local water source at a shallow arctic lake located on Alaska's Coastal Plain, 50 km south of Prudhoe Bay, USA. A 130 m long and 3 m tall snow fence was placed next to a lake for the winters of 2009–2010 and 2010–2011. This snow fence retained large quantities of snow transported by wind during the winter, forming a sizable snowdrift. In this region of the Arctic, lakes receive most of their water from seasonal snowpack melt in late May and early June, with subsequent drying of the lakes after snowmelt. The artificially created snowdrift at the experimental lake contributed meltwater until early July during both summers of the study period and effectively offset summer lake drying. At summer's end, the additional lake volume from the fence snowdrift was 21% and 29% in 2010 and 2011, respectively. While techniques such as snow fences can be used to address existing needs for enhanced water supplies, they can also serve as a tool for assessing lake response to a future climate scenario in which summers are warmer and both evaporation and winter precipitation increase.

Keywords

Arctic, water supply, lake, snowdrift, snow fence

1. Introduction

Water availability in the Arctic shifts dramatically between the seasons, with an abundance in spring and scarce supply in winter. Constant local water supply in the polar regions can be difficult to maintain because of the long periods of continuous freezing temperatures and the presence of snow cover and permafrost (McFadden and Collins, 1978; Slaughter et al., 1975). Prolonged freezing air temperatures last from October to May in the Alaskan Arctic (Olsson et al., 2002). During this time, most of the ground surface is covered by snow and surface waters are covered by ice. Air temperatures might rise above freezing and affect snowpack conditions, but before May, energy is insufficient to overcome the snowpack's cold content and generate surface water runoff. In addition to the cold climate, large parts of the Arctic receive little precipitation, less than 250 mm annually, which makes this region a polar desert (Woo and Young, 2014). About half of the annual precipitation is stored as seasonal snow cover and then released to the drainage network during snowmelt, typically in middle May through early June. These environmental factors cause challenges that affect the local water supply, including: (1) short open-water season in rivers and lakes, (2) low annual precipitation, (3) limited access to groundwater resources constrained by continuous permafrost, and (4) high operating costs of water distribution systems (see Slaughter et al., 1975).

Arctic communities and industry largely depend on the local water supply for domestic purposes, maintenance of existing infrastructure, and oil and gas exploration and industry activities. This project was funded to address environmental questions associated with allocation of water resources for the construction of ice roads and ice pads. An ice pad is a temporary ice structure built in wintertime for the purpose of having a work area for exploratory drilling on the tundra with minimal environmental impact. Ice roads provide the means by which supplies,

materials, equipment, and machinery are transported to remote exploration sites, otherwise inaccessible by conventional roads. A traditional water source for ice road construction is water withdrawal from lakes (Hinzman et al., 2006), as local streams and rivers have very low or a complete lack of winter flow.

The distinctive hydrological regime of Arctic lakes, caused by the presence of seasonal ice cover and permafrost, exerts an influence on lake water availability in winter (Arp et al., 2012; Duguay et al., 2003). Lakes are covered with ice from October to June, and often no water recharge of lakes occurs until snowmelt in early June. The shallow active layer has limited storage, which minimizes suprapermafrost groundwater contributions to lake systems. Most lakes in permafrost environments have a thaw bulb (talik) of unfrozen ground directly beneath them. Subpermafrost groundwater can contribute water to the larger lakes via open taliks that penetrate permafrost completely, like Teshekpuk Lake (Kane et al., 2013). There is no subpermafrost water contribution to the majority of smaller lakes, as they have closed taliks. After snowmelt, water volumes in the lakes decrease throughout the summer, when water loss due to evaporation is considerably greater than water gained from rainfall (Bowling et al., 2003; Mendez et al., 1998). This balance switches in August, when air temperature drops, evaporation decreases, and rain (or snow) is more likely to occur. Some of the summer surface storage deficit in the active layer and surface water bodies (lakes, ponds, wetlands) is recharged during this time (Bowling et al., 2003). However, if the surface storage deficit is not replenished (for example, precipitation in the fall is low and near-surface soils are dry), lake recharge is directly affected, and water availability for the following winter is reduced. For example, an extremely dry summer and fall occurred in 2007. During that year, the Kuparuk River at Deadhorse, Alaska had its lowest end-of-summer flow since 1971, when the U.S. Geological Survey initiated

streamflow gauging at this site (Kane et al., 2008). The impact of the 2007 drought on the surface storage deficit was evident during the 2008 snowmelt runoff and annual water yield (Arp et al., 2012).

One of the ways to address the consequences of drought and drought-like conditions is to augment a lake's water supply with snow fences and snowdrifts (i.e., Jairell and Schmidt, 1990; Tabler, 1968). In watershed hydrology, natural snowdrifts have a profound effect on energy and mass balance at certain spatial scales (Clark et al., 2011; Davison et al., 2006; Grayson and Blöschl, 2001; Woo and Young, 2014; Young et al., 2013). Over time, snowdrifts and snow fences have proven to be effective tools for harvesting water. More than a century ago, an Alaskan newspaper—*Seward Weekly Gateway*—described the successful use of snow fences by Seward miners in 1909 (Seward Weekly Gateway, 1909). In the 1970s and 1980s, both in Wyoming and in some Alaska Arctic villages, significant progress was made in increasing local water supplies by using snow fences (Jairell and Schmidt, 1990; McFadden and Collins, 1978; Sturges, 1992; Tabler, 1968). Recently, snow fences were used in Afghanistan to increase water yield for irrigation (Thompson et al., 2009).

Here, we examine both naturally occurring and artificially created snowdrifts to quantitatively assess the response of a lake to seasonal precipitation storage within the lake's watershed. This study is twofold: the first part focuses on the characteristics of snowdrifts at selected locations, precipitation storage capacity of snowdrifts, and rate of snowdrift melt; the second part quantifies the change in lake water levels and volumes during the open-water season in response to water contribution from an artificially created snowdrift.

2. Study Area

Lakes, ponds, and wetlands are extensively distributed throughout the Arctic. Lakes account for 15–40% of the circumpolar Arctic Coastal Plain (ACP) (Duguay et al., 2003) and for at least 25% of Alaska's ACP (Arp et al., 2012; Sturm and Liston, 2003). The majority of Alaska's ACP lakes are shallow with a low ratio of lake volume to lake surface area (Arp and Jones, 2009; Sellmann et al., 1975).

In selecting a research site, we focused on lakes accessible from the Dalton Highway (Figure 1). Beyond the highway's terminus at Deadhorse are private roads owned by oil companies, with use restricted to authorized vehicles only. We selected two lakes 30 miles south of Deadhorse near Franklin Bluffs (Figure 1). One lake is a gravel pit (also referred to as the experimental lake); it serves as a water source for maintaining the nearby Dalton Highway. The second lake is a natural lake (referred to as the control lake), representative of shallow thermokarst ACP lakes.

Low-gradient terrain prevails at the study area. Generally, the elevation gradually increases from the coastline toward the foothills of the northern Brooks Range, which borders Alaska's ACP to the south and east. The flat coastal landscape features some steep cliffs at riverbanks and lakes that can be as high as 5 to 10 m. Higher terrain occurs at White Hills and Franklin Bluffs (Figure 1). The entire region is underlain with continuous permafrost that reaches a maximum thickness of 600 m near Prudhoe Bay (Osterkamp and Payne, 1981). The active layer is typically 50 to 90 cm and varies considerably due to vegetation, soil moisture, soil type, aspect, slope, etc., with extensive surficial organic soils overlying mineral soils (Slaughter and Kane, 1979). This region has low-growing tundra vegetation. Sedge tussocks, dwarf (< 0.5 m) deciduous shrubs, and occasional large (2–7 m high) deciduous shrubs prevail along floodplains, water tracks, and other drainages (Tape et al., 2006).

Figure 1. Study area. Map at upper left shows the study lakes and Alaska's ACP near the Dalton Highway and Deadhorse. Upper right panel zooms in to the experimental lake, snow fence, and location of the 10 m tower. Bottom panels are photographs of the experimental and control lakes and view of Franklin Bluffs.

3. Methods and Data

A suite of field measurements, including snow surveys, weather, and hydrologic data, was collected at the experimental lake for one spring and one summer before the snow fence was installed. These measurements, taken in spring and summer 2009, represent baseline conditions.

The same suite of measurements was collected for two more years after a 130 m long and 3 m tall snow fence was placed next to the experimental lake in September 2009 (Figure 2). Data collected at the control lake served as a reference to the natural hydrological regime during this two-year experiment. Overall, field data collection lasted from April 2009 to September 2011.

Figure 2. Sketch showing the experimental lake setup during the baseline year before installing a snow fence (A) and during the two-year experiment after the snow fence was installed in September 2009 (B). The sketch is not to scale.

3.1 Snow survey

Detailed snow depth information was collected for each lake and from nearby tundra using a MagnaProbe (Sturm and Holmgren, 1999). The sampling extent covered both the lake surface and the surrounding tundra because snow properties differ depending on the underlying material (lake ice versus tundra vegetation). The sampling strategy included detailed snow depth measurements approximately every 2 m in a rectangular grid, transects, and random patterns.

The automated snow depth probe (MagnaProbe) recorded both the snow depth and the geographical position of the each snow depth measurement. Collected snow depth measurements were interpolated using the nearest-neighbor technique in a regular grid with a 10 m cell size. The resulting gridded dataset was then used for snow depth and snow volume analysis. Snow volume (Ψ_s) was calculated using

$$\Psi_s = \sum_{i=1}^n h_{s_i} A_i \quad \text{Eq. 1}$$

where h_{s_i} is the snow depth at i th grid cell (m), and A_i is the area of the grid cell (m²). Periodic snow density measurements on the lake surface and on the tundra accompanied the snow depth measurements (Table 1). Snow density was measured with a fiberglass tube (Adirondack) that had an inside area of 35.7 cm² and was equipped with metal teeth on the lower end to cut through dense snow layers.

Table 1. Summary of the measurements.

Time period	Type of measurement	Number of measurements		
		Experimental Lake	Natural Lake	Snow Fence
24–25 April 2009	Snow depth	1,359	100	n/a
	Bulk density	15	10	n/a
22 April–3 June 2010	Snow depth	5,167	9,812	419
	Bulk density	50	15	2
	Drift survey	n/a	n/a	2
	Snow pit	n/a	n/a	2
14 April–5 June 2011	Snow depth	1,754	1,098	297
	Bulk density	48	10	2
	Drift survey	n/a	n/a	2
	Snow pit	n/a	n/a	2

To evaluate seasonal snow accumulation during the three years of this experiment in terms of long-term average maximum and minimum snow water equivalent (SWE), we used the snow-research dataset collected by UAF/WERC faculty, staff, and students from 1999 through 2014 (Stuefer et al., 2013). This longer record includes SWE calculated from 50 snow depths and 5 snow densities using Eq. 2.

$$SWE = h_s \frac{\rho_s}{\rho_w} \left(\frac{10 \text{ mm}}{1 \text{ cm}} \right) \quad \text{Eq. 2}$$

where SWE is snow water equivalent (mm), h_s is the average snow depth (cm), ρ_s is the average bulk snow density (kg/m^3), and ρ_w is water density (1000 kg/m^3).

3.2 Snow fence and snowdrift

A Wyoming-type snow fence 3 m tall and 130 m long was constructed in September 2009 (Tabler, 2003). The snow fence was large enough to create a snowdrift that would capture a volume of snow comparable to the available lake storage. We used empirical relationships between snow transport, winter precipitation, and fetch published in Tabler (2003) to verify that the desired size of the snowdrift was achievable in the study environment. In addition to the snow fence, we installed instrumented towers on the downwind side of the fence at 10, 16, and 34 m to monitor snowdrift growth (Figure 3A, 3B). Snow drift growth was monitored in a line perpendicular to the fence, and the snow depth profile along this line is referred to as a central profile (Figure 3C).

Figure 3. Schematic diagram (A) and photograph (B) showing the snow fence setup. Bottom panel (C) illustrates geometrical shapes and corresponding areas used to calculate daily snowdrift melt. Symbols h_1 , h_2 , and h_3 represent snow depths measured by the SR50 sensors. Symbols d_1 , d_2 , d_3 , and d_4 represent horizontal distance constrained by the SR50 sensors.

Each tower was equipped with a sonic depth sensor SR50 that measured the distance from the sensor to the surface (ground or vegetation surface in summer and snow surface in winter). These three towers were connected to a fourth tower, which was placed next to the fence. This fourth “snow fence” tower was equipped with the data logger, power supply, solar panel, camera, antenna, radio, and relative humidity (RH) and air temperature (T_{air}) sensors. Snowdrift surface elevations were measured at the winter’s end to estimate snowdrift volume. We surveyed the surface elevation using a high precision Differential Global Positioning System (DGPS; Trimble

5700) and post correction technique. Horizontal and vertical positions recorded by this system have an accuracy of ± 2 cm (Sturm and Stuefer, 2013). The snow pit was excavated at the crest of the snowdrift (near snow depth sensor 1) to evaluate the snow stratigraphy. A continuous snow density profile was measured from the top to the bottom of the snowdrift every 3 cm (i.e., the height of the density cutter). Bulk snowdrift density was calculated by averaging the measured snow density profile.

3.3 Weather data

The NOAA National Weather Service and NRCS SNOTEL, as well as other agencies collecting hydrometeorological data, have limited network coverage over this region (Kane and Stuefer, 2015). Severe climate conditions, an abundance of wildlife, and lack of access prevent regular data collection by most agencies. Stations here require high-cost maintenance to provide quality data—a source of invaluable information for engineers, scientists, and local communities. Independent automated data collection and collaboration with other ongoing research projects are common practices when conducting research in this area.

We used a weather station at the Franklin Bluffs site that had been established and maintained by UAF/WERC under a Department of Energy research project in 1988. This weather tower (10 m tall), located 500 m northeast of the experimental lake, is equipped with air temperature, relative humidity, sonic snow depth, and wind speed and direction sensors (Figure 1). Air temperature, wind speed, and relative humidity were measured at 1 m, 3 m, and 10 m heights above the ground surface. The rainfall gauge and an evaporation pan were located within 5 m of the weather tower. Rainfall data were collected with a tipping bucket rain gauge equipped with an Alter shield.

3.4 Water level and volume

In both the experimental and control lakes, water levels were measured with vented pressure transducers. The experimental lake had continuous water level records from 9 June 2009 to 31 August 2011. The control lake had water level records from summer 2009 and 2010. The water level–volume relationship was established to calculate lake volume from water level measurements. We conducted a bathymetry survey on 28 July 2009. The experimental lake water surface area was measured at 21,000 m², and water volume was calculated at 21,700 m³. Some parts of the experimental lake were very shallow 0.4–0.6 m, but in some places, water depth was more than 2 m, meaning that the deepest part of the lake remained unfrozen in winter. The average depth of the experimental lake was 1.1 m. The control lake water surface area and volume were measured at 390,000 m² and 479,700 m³, respectively. The control lake is shallower than the experimental lake, with a maximum depth of 1.6 m and an average depth of 1.2 m; it freezes completely during winter. Bathymetry data were processed with geographical informational system software (3D Analyst, ArcMap) to interpolate and extrapolate volumes for the range of water levels and to obtain the water level–volume relationship (Figure 4).

Figure 4. Relationship between lake water volume and water level for the observed range in water levels.

During each summer, water from the experimental lake is used by the Alaska Department of Transportation and Public Facilities (ADOT&PF) mostly for dust control on the Dalton Highway. The volume of water taken from the experimental lake during years of this study was 2756 m³ in summer 2009, 7532 m³ in summer 2010 and 2271 m³ in summer 2011.

3.5 Snowdrift melt

The rate of snowdrift melt was calculated directly from the continuous measurements of snow depth at three locations along the central snowdrift profile. The following steps were taken to estimate the entire snowdrift volume from these point measurements. First, we used three snow depths to calculate snow volume per meter of fence at the central profile (Figure 3C). To make this calculation, the snowdrift shape was approximated with four simple geometric shapes, constrained by the location of each SR50 snow depth sensor, as shown by the schematic diagram in Figure 3C. One assumption for this model was that the snowdrift shape satisfies geometric similitude as it melts; that is, ratios h_1/d_1 and h_3/d_4 remain constant at each time step during snowmelt calculations (Figure 3C). The sum of these four areas gives the volume of the snowdrift per meter of fence at the central profile at a given time. Second, we multiplied this unit snow volume from the central profile by 100 m to obtain the entire snowdrift volume. The selected length (100 m) is 30 m less than the actual snow fence length (130 m) to account for reduced snow-capturing capacity at both ends of the snow fence (Tabler, 1980).

Snow water equivalent was calculated using Eq. 2, where the average snow depth was obtained by dividing snowdrift volume by snowdrift area. Bulk snow density for SWE calculations during snowmelt was taken from the last snow pit measurements on 1 June 2010 ($481 \text{ kg/m}^3 \pm 62 \text{ kg/m}^3$) and 5 June 2011 ($519 \text{ kg/m}^3 \pm 51 \text{ kg/m}^3$). Snow density was assumed constant for the rest of the snowmelt period, and corrections for snow densification due to melt

was not performed, since we did not have measurements to validate such a density adjustment.

The difference between the modeled and measured snowdrift volume ranged from 1% to 6%

(Table 2).

Table 2. Comparison of snowdrift volume (V) calculated from the snow surface elevation survey with DGPS (observed) and from the sonic snow depth sensor SR50 records (modeled).

	24 April 2010		30 May 2010		16 April 2011		5 June 2011	
	V_{obs}	V_{mod}	V_{obs}	V_{mod}	V_{obs}	V_{mod}	V_{obs}	V_{mod}
Snow volume, m^3	21,378	20,650	19,315	19,118	16,152	15,893	10,467	9,820
Snow density, kg/m^3	414		481		435		519	
Water volume, m^3	8,850	8,531	9,291	9,196	7,026	6,913	5,432	5,097

3.6 Evaporation

Water loss in summer occurs due to evaporation from both the surface of the snowdrift and the water surface of the lake. This section describes two methods used to quantify evaporation rate. The rate of evaporation from the snowdrift (E) or condensation onto the snowdrift (C) was calculated from latent heat flux (Dingman, 2008):

$$E \text{ or } C = \frac{LE}{\lambda_v \times \rho_w} \left(\frac{1000 \text{ mm}}{1 \text{ m}} \right) \quad \text{Eq. 3}$$

where E or C is the rate of evaporation or condensation (mm/hr), LE is the latent heat exchange between the snowdrift surface and the overlying air ($\text{kJ}/\text{m}^2/\text{hr}$), ρ_w is the density of water ($1000 \text{ kg}/\text{m}^3$), and λ_v is the latent heat of vaporization ($2496 \text{ kJ}/\text{kg}$) required to transform liquid water to vapor. Latent heat was calculated following Price and Dunne (1976), as summarized in Appendix 1. Latent heat flux calculations were conducted only during the snowdrift melt period from 1 June to 7 July 2010 and from 1 June to 4 July 2011, using hourly relative humidity, air temperature, wind speed, and atmospheric pressure measurements. The vapor pressure of the overlying air was calculated as described in Dingman (2008), using hourly relative humidity and

air temperature measured directly above the snowdrift (see Figure 3A for the location of this sensor).

Common assumptions employed for the latent heat flux calculations during the snowmelt period (1 June to 7 July) included (1) stable atmospheric conditions, (2) logarithmic profiles for wind speed distribution, (3) snowdrift surface temperature T_s equals 0°C, and (4) vapor pressure at the snow surface equals saturation vapor pressure (6.11 mbar at $T_s = 0^\circ\text{C}$).

Evaporation from the water surface of the lake was estimated using a different technique, since not enough data were available to calculate latent heat flux at the lake surface. A standard Class A evaporation pan (used by the U.S. National Weather Service) was used to measure potential evaporation during the summer season. Seasonal lake evaporation was determined by adjusting potential evaporation to account for advection and heat-storage effects in a given lake (Dingman, 2008):

$$E = k [P - (V_2 - V_1)] , \quad \text{Eq. 7}$$

where E is lake evaporation during the chosen period (mm), V_1 and V_2 are the water depths in the evaporation pan at the beginning and end of this period (mm), P is precipitation during that time (mm), and k is the ratio of lake evaporation to pan evaporation or the pan coefficient (dimensionless). Clearly, an evaporation pan differs from a lake, in that it has far less heat-storage capacity, lacks surface or groundwater inputs or outputs, and has sides exposed to the air and sun and bottom exposed to air circulation (Dingman, 2008). These differences significantly affect the energy balance, elevating the warm season average water temperature and vapor pressure of the pan's water surface relative to that of a nearby lake (Dingman, 2008). The local pan evaporation coefficient, 0.85, was reported for the lakes located in the vicinity of Franklin Bluffs (Kane and Carlson, 1973; Mendez et al., 1998) and applied in this study.

4. Results

4.1 Precipitation

Between 2000 and 2014, the annual precipitation at research site varied widely from 75 mm in 2007 to 281 mm in 2003 (Table 3). On average in this location, 62% of the annual precipitation accumulates in the seasonal snowpack. Over the period of record the percentage of snow precipitation in annual precipitation varied from 38% to 88% (Table 3). The cold season referred to in Table 3 extends from September to May, with the exact dates refined by air temperature observations. The warm season starts (and cold season ends) when the daily average air temperature remains above freezing for several days, usually in the second half of May. The warm season ends (and cold season starts) when the daily average air temperature stays consistently below freezing, usually in late September.

Table 3. Summary of precipitation at the Franklin Bluffs weather station. End-of-winter SWE is used for cold season precipitation.

Water Year (1 Oct to 30 Sep)	Cold Season		Warm Season		Annual mm
	mm	% of annual	mm	% of annual	
1999 – 2000	145	60	98	40	243
2000 – 2001	134	59	93	41	227
2001 – 2002	85	38	139	62	224
2002 – 2003	195	69	86	31	281
2003 – 2004	91	45	113	55	204
2004 – 2005	115	69	52	31	167
2005 – 2006	102	50	100	50	202
2006 – 2007	66	88	9	12	75
2007 – 2008	97	57	74	43	171
2008 – 2009	160	64	89	36	249
2009 – 2010	117	65	63	35	180
2010 – 2011	136	72	54	28	190
2011 – 2012	139	59	96	41	235
2012 – 2013	72	40	110	60	182
Average	118	62	84	38	202

The purpose of this section is to compare winter, summer, and annual precipitation in 2009, 2010, and 2011 with each other and with the historic record. The SWE measured in late April often serves as a metric for total winter precipitation (less sublimation), because precipitation gauges tend to significantly undercatch snowfall in the Arctic (Goodison et al., 1998).

Fourteen years of end-of-winter snow survey data show that the Franklin Bluff site accumulates on average 118 mm of SWE in the seasonal snowpack (Figure 5). In comparison with the 14-year average, end-of-winter SWE was significantly above average in 2009 (SWE = 160 mm), close to average in 2010 (SWE = 117 mm), and above average in 2011 (SWE = 136 mm).

Figure 5. Cold season and warm season precipitation. Top panel shows 14-year record of end-of-winter SWE on the tundra near the Franklin Bluffs meteorological station, 2000–2013. Each bar represents SWE calculated from 50 snow depths and 5 snow densities. The bottom panel shows a family of curves representing cumulative summer rainfall at Franklin Bluffs from 1988

to 2013 (not adjusted for wind undercatch). Black lines represent cumulative summer rainfall during study years.

Cumulative summer rainfall (1988–2013) ranged from 9 mm in 2007 to 132 mm in 2002 (Figure 5). During the years of the snow fence experiment, the lake received 63 mm of rainfall in summer 2010 and 54 mm of rainfall in summer 2011 (Table 3). During summer 2009 (baseline summer), the lake received 89 mm of rainfall. Thus, summer 2009 was the wettest year during three study years in terms of both seasonal snow accumulation and rainfall.

4.2 Snow fence drift

Only part of the cold season precipitation stays at the tundra surface where it falls; the rest is moved by wind to a different location. During the first winter (2009–2010), the snow fence captured 21,378 m³ of snow by 24 April 2010 (Table 2). During the second winter (2010–2011), the end-of-winter snow accumulation in the drift was lower—16,152 m³ of snow on 16 April 2011—mainly because the height of the fence was reduced to 2.6 m as result of maintenance.

Continuous records of snowdrift growth with the SR50 sensors have allowed us to identify the number of snow deposition events that formed the snowdrift. Seven snow deposition events that built the snowdrift are visible during winter 2009–2010 (Figure 6). Four snow deposition events shaped the snowdrift during winter 2010–2011 (Figure 6). The snowdrift was 3 m deep as early as 4 November 2010. Snow deposition on the snowdrift was recorded when hourly averaged wind speed at the 10 m height exceeded 8 m/s; significant snowdrift growth (approximately 1 m during a couple of hours) corresponded to wind speeds above 17 m/s. Snow deposition rates varied from 1.3 to 3.9 m³/m/hr (cubic meters of snow per 1 m of fence per hour). This methodology is described in Sturm and Stuefer (2013).

Figure 6. Snowdrift evolution as measured by snow depth sensors (see locations in Figure 3). Numbers indicate snow deposition events that formed the snowdrift.

On Alaska's ACP, where the snowpack is generally cold and dry, the wind regime has primary control over snowdrift size (Sturm and Stuefer, 2013). The predominant wind direction at the study lakes is northeast, with occasional strong wind from the southwest (see Figure 7). During 2009, 2010, and 2011, southwest winds were more pronounced during December, January, and February, and they were barely noticeable during other months. Strong winds were most likely to occur during December, January, and February (76 hours total) and were less likely to occur during June, July, and August (only 2 hours total). The average hourly wind speed was 4.5 to 4.7 m/s for most of the months, and was slightly higher—5.2 m/s—during the mid-winter quarter (December, January, and February). Wind speed is reported at a 10 m height unless a different height is specified.

Figure 7. Hourly averaged wind speed at 10 m height, plotted as relative frequency of occurrence within 10° arc segments. These data represent three years: 2009, 2010, and 2011.

4.3 Snow distribution on the lake

Any topographic feature, tall-growing vegetation or infrastructure in the pathway of blowing snow results in snowdrifts of various sizes and shapes. Both study lakes have natural snowdrifts along the shoreline. These snowdrifts are located along the southwest part of the lakes at cut banks (Figure 8). To isolate and quantify the effect of natural snowdrifts, lake snow cover was separated into two classes: shoreline snowdrift (deep snow in Figure 8) and lake snow cover without shoreline snowdrift (shown as crosshatched area in Figure 8).

Figure 8. Natural snow distribution on the lakes. Purple indicates location of the shoreline snowdrift. Area with uniform snow distribution is crosshatched. Shoreline snowdrift covers a small part of the surface area, but accumulates half of the lake's snow.

Shoreline snowdrift represented a relatively small percentage of lake area, 20%, yet it accumulated almost half (48%) of the lake's snow by the end of winter in 2010 (Table 4).

Table 4. Lake snow cover characteristics.

	Surface Area		Snow Depth (cm)		Snow Volume	
	km ²	%	mean	st. dev.	(×10 ³) m ³	%
15 April 2011 (natural lake)						
Entire lake	0.391	100	39	20.3	153,510	100
Lake snow without drift	0.307	78	31	4.0	96,594	63
Shoreline drift on the lake	0.084	22	68	28.8	56,917	37
24 April 2010 (natural lake)						
Entire lake	0.391	100	18	17.2	70,678	100
Lake snow without drift	0.314	80	12	4.2	36,402	52
Shoreline drift on the lake	0.077	20	45	23.2	34,276	48
24 April 2009 (experimental lake)						
Entire lake	0.0210	100	55	18.1	11,522	100
Lake snow without drift	0.0156	74	46	10.1	7,222	63
Shoreline drift on the lake	0.0054	26	80	12.4	4,299	37

4.4 Snowmelt

Seasonal tundra snowpack commonly disappears within the 2–3-week-long ablation period that occurs in May or early June. The ablation period of the artificial snowdrift lasted a month longer than the snowmelt of seasonal snowpack on the surrounding tundra (Figure 6 and Figure 9).

As seasonal tundra snowpack melts in May, snowmelt pools in the low-centered polygons, depressions, and polygon troughs. The soils are still frozen at that time. Available surface water storage usually fills within a few days, and snowmelt water spills to the lake through a temporary channel (see photo in Figure 9). This meltwater turns a completely snow-covered lake into a patchy mix of water, ice, and snow within less than a day. Its effect is similar to a rain-on-snow event, transferring heat from the warmer tundra water to the colder lake snowpack and releasing latent heat of fusion, if/when this water refreezes. Most of the water contained in the lake snowpack, including the natural shoreline snowdrift, contributes directly to the lake water balance.

Figure 9. Depletion curve of snowdrift and natural snowpack in 2010 and in 2011. Photographs show tundra snowmelt entering the lake surface, transforming the snow-covered lake surface into a patchy mix of water, ice, and snow.

The snowdrift at the snow fence started melting later, in the beginning of June, as it took longer to overcome the cold content of the deeper snowpack (Figure 6 and Figure 9). The cold content of snowpack is the amount of energy required to raise its average temperature to the melting point (Dingman, 2008). Snowdrift ablation took more than one month—32 days in 2010 and 34 days in 2011. During this time, most meltwater flowed to the lake (Figure 9), while the loss of snowdrift mass to evaporation was negligible (see Section 4.5 for more details). Snowdrift melt contributed water to the lake long after the natural shoreline snowdrift and tundra seasonal snowpack had disappeared. Given the Arctic’s short summers, with the warmest months in June and July, this additional snowmelt water yield offsets drying of the lake for the entire month of June.

4.5 Evaporation

The rate of evaporation from the snowdrift was calculated from latent heat flux Eq.3. The snowdrift lost very little mass due to evaporation in the summer months. Cumulative evaporation from the snowdrift was 1.5 mm from 1 June to 7 July 2010 and 15 mm from 1 June to 7 July

2011. Saturation vapor pressure at the snow surface ($e_s = 6.11$ mb at $T = 0^\circ\text{C}$) was consistently lower than the vapor pressure above the snowdrift (Figure 10). This difference in vapor pressure between snowpack and overlying air favored condensation onto the snowdrift surface (Figure 10), resulting in 71 mm of condensation from 1 June to 7 July 2010 and 56 mm of condensation from 1 June to 4 July 2011.

Figure 10. Evaporation and condensation during melt of the snowdrift. Vapor pressure gradient represents the difference between vapor pressure of the air above the snowdrift (e_a) and saturation vapor pressure at the snow surface (e_s).

High evaporation rates promote drying of the lakes. The difference between rainfall and evaporation (P-E) for the experimental lake equaled -96 mm in 2009, -134 mm in 2010, and -89 mm in 2011. This difference was reported for the same period from 23 June to 26 August.

4.6 Lake volume without the artificial snowdrift

Baseline hydrologic observations showed that water levels in both lakes were rising in June in response to snowmelt. This observation is well documented (i.e. Bowling et al., 2003; McNamara et al., 1998, Rovanešek et al., 1996). In summer, when water loss due to evaporation is considerably greater than water gain due to rainfall, lake levels gradually drop. An example of lake level fluctuation during a dry summer is shown in Figure 11.

Figure 11. Water levels observed in the control lake in summer 2010.

Results of the weather data analysis summarized in the previous sections showed that weather conditions (SWE, rainfall, and evaporation) were conducive to relative high lake water levels in summer 2009. The first year of the experiment (2010) represented a dry summer during which the natural lake water levels dropped (Figure 11). Even though water level data are not available for the control lake in 2011, weather data indicate that summer 2011 was similar to summer 2009 in terms of P-E.

4.7 Effect of snowdrift on experimental lake volumes

We evaluated the change in lake volumes during the two-year experiment by comparing them with the lake volumes in summer 2009 before installation of the snow fence. This approach resulted in conservative estimates in summer 2010, since wetter summer conditions in 2009 were conducive to higher than normal lake volumes.

The experimental lake had water withdrawals by ADOT&PF for maintenance of the Dalton Highway (Figure 12 and Table 1). Once this pumped water was added to the observed water volumes, we could obtain reconstructed lake water volumes that reflect lake behavior as if no water had been withdrawn from the lake. These reconstructed lake water volumes were used in the analysis to quantify the effect of the snowdrift (Figure 13).

Figure 12. Observed water volume in experimental lake during summers of 2009, 2010, and 2011. Thick lines indicate water pumping by ADOT&PF for maintenance of the Dalton Highway.

Figure 13. Reconstructed water volume in the experimental lake during summers of 2009, 2010, and 2011.

Lake volumes remained elevated through the summers of 2010 and 2011 due to snowmelt released from the snowdrift (Figure 13). By the time the snowdrift had melted in early July, lake volumes had increased by 23–27% in comparison with the lake volume on 4 July 2009, the baseline summer (Table 5). The end-of-summer increase in lake volumes was 21–29% in comparison with the 2009 lake volumes (Table 5). Water balance summary for each summer was presented in Stuefer and Kane (2013).

Table 5. Experimental lake water volumes (m³).

	2009	2010	2011
Lake volume on 4 July	26,129	31,867	33,154
Increase in water volume from 2009 (on 4 July)		5,981 (23%)	7,025 (27%)
Water volume on 26 August	23,914	28,858	30,764
End-of-summer increase in water volume from 2009 (on 26 August)		4,945 (21%)	6,851 (29%)

A comparison of snowdrift water volumes (Table 2) and total increase in lake volumes (Table 5) indicates a nonlinear response; that is, a smaller drift in 2011, but more water in the lake. This is partly attributed to the drier summer in 2010 and the wetter summer in 2011, but it

may partly be attributed to the influence of other components of the lake's hydrologic system, such as changes in both surface and subsurface storage between the years of the experiment.

5. Discussion and Conclusions

Tabler (1968), and later Slaughter et al. (1975), formulated three general environmental criteria for successfully capturing snow with snow fences and using the resulting snowdrifts as a local water supply. To paraphrase Tabler (1968), (1) snow must be a significant source of water yield in the area; (2) weather conditions (wind, precipitation, and air temperature) must be conducive to snowdrifting; and (3) snow retention in the area can be increased significantly. To these criteria, we can add that low permeability of underlying soils and proper snowdrift placement are required for effective delivery of snowdrift water to the drainage network (Jairell and Schmidt, 1990; Sturges, 1989). In a low-gradient terrain, such as that studied here, a snow fence should be placed within the distance of the expected snowdrift length ($20H$, Figure 2). The approach of using snow retention to augment a lake's freshwater supply can be applied at any location that satisfies the criteria listed above.

These criteria were met at Alaska's ACP in the past; we have shown that they work well in present conditions; and these criteria may play an important role in a warmer future climate. The present experiment with an artificially built snowdrift has proven useful for increasing water yield during the year. This experiment can also serve as a hypothetical example of lake response to increased winter precipitation. Future climate projections in this part of the Arctic indicate an increase in air temperature, evapotranspiration (ET), and winter precipitation (Barnett et al., 2005; Hinzman et al., 2013). A projected increase in ET is likely to decrease the summer (P-ET); and that, in turn, will promote further drying of the surface storage, including lakes, ponds, and wetlands. In winter, however, a projected increase in precipitation (Krasting et al., 2013) can

contribute to full-capacity snowdrift growth, assuming a continuing abundant wind supply and air temperatures low enough to maintain natural properties of tundra snowpack. While projected increase in ET is likely to promote drying of the lakes, projected increase in winter precipitation might offset the P-E deficit during following summer.

The results of this experiment have shown that large snowdrifts next to a lake are capable of effectively offsetting the increase in P-E deficit, thus preventing drying of lakes in the summer. Bowling et al. (2003) showed that summer evaporation and drainage result in a seasonal reduction of surface water extent ranging from 58–73% in this part of the Arctic. We showed in the present study that snowdrift water helps reduce summer lake drying and increases usable water. This increase in water availability is attributed mainly to the mass of the snowdrift, its month-long extended melt, and the local “recycling” of water vapor from the overlying air during the summer months. As the vapor pressure in overlying warm air often exceeds the saturation vapor pressure at the snow surface at 0°C, water vapor condenses onto the snowdrift surface, generating additional water.

There is a limit to lake size where natural and artificial snowdrifts can substantially increase water volume. The described technique can be effective at lakes of small and medium size. Following Sturm and Liston (2003), we suggest that the impact of the natural snowdrifts on the winter mass balance of the lake can be noticeable for a lake with an area less than 400 ha. Artificial snowdrifts can be effective in producing a 21–29% increase in usable water amount when the amount of snow collected by the fence is comparable to the volume of the lake. In our experiment, drift snow volume constituted 82% of the baseline lake volume in 2010 and 62% in 2011.

Snowdrifts regulate water and energy exchanges within the hydrologic system at local scales. The local spatial scale represents distances of several to hundreds of meters up to several kilometers and are associated with uneven terrain, hillslopes, ridges and/or vegetation (Liston, 2004; Male and Gray, 1981). As science and technology progress, modeling of hydrologic processes requires resolving hydrologic processes at these smaller scales. Data presented in this work can be used to improve representation of snow-related hydrological processes on the local scale, where the presence of snowdrifts is pronounced.

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6. Appendix 1

Latent heat was calculated following Price and Dunne (1976):

$$LE = \lambda_v \rho_a D_e \frac{0.622}{P} (e_a - e_s^*) \quad \text{Eq. A1}$$

where ρ_a is the density of air (1.15 kg/m^3), D_e is the exchange coefficient for moisture (m/hr), P is the atmospheric pressure (mb), e_a is the vapor pressure of the surrounding air calculated from the relative humidity measurements directly above the snowdrift (mb), and e_s^* is the vapor pressure at the snow surface (mb), which is assumed to be the saturation vapor pressure at the snow surface temperature. Evaporation occurs when vapor pressure at the surface is greater than the vapor pressure of overlying air ($LE < 0$); otherwise, water vapor condenses onto the snow surface ($LE > 0$).

The exchange coefficient D_e for neutral atmospheric stability conditions is given by the analogy with the transfer of momentum (Price and Dunne, 1976):

$$D_e = \frac{k^2 u_z}{[\ln((z-h)/z_0)]^2} \left(\frac{3600 \text{ s}}{1 \text{ hr}} \right) \quad \text{Eq. A2}$$

where k is von Karman constant (0.4), u_z is the wind speed at z height (10 m) above ground (m/s), h is the depth of snow (m), z_0 is the roughness length (0.001 m or 1 mm), and the last term is the unit conversion factor from m/s to m/hr.

Stability conditions (stable or unstable) of the air overlying the snow surface affect the rate of latent heat exchange. During snowdrift melt, air in contact with the colder snow surface tends to cool, becoming denser with a decrease in the rate of turbulent exchange. This situation represents stable conditions, and $(D_e)_s$ can be described with the following equation (Price and Dunne, 1976):

$$(D_e)_s = \frac{D_e}{1+10Ri} = \frac{D_e}{1+10\{[g(z-h)/(T_a-T_s)]/[(T_a+273.2)(u_z)^2]\}} \quad \text{Eq. A3}$$

where $(D_e)_s$ is the exchange coefficient under stable atmospheric conditions (m/hr), Ri is the Richardson number following the formulation in Kane et al. (1997), g is the gravitational constant (9.81 m/s^2), T_a is the absolute air temperature (K), and T_s is the absolute snow surface temperature (K). Values of Richardson number greater than 0.2 represent conditions where no convection would occur (DeWalle and Rango, 2008).

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