The role of winter sublimation in the Arctic moisture budget*

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Abstract In the Arctic, the simplest way to describe the winter surface moisture budget (in the absence of any net horizontal transport) is: snow-water-equivalent depth on the ground (*D*) equals precipitation (*P*) minus sublimation (*S*). *D*, *P* and *S* are the most fundamental components of the winter arctic hydrologic cycle and understanding them is essential to understanding arctic moisture-related processes. Unfortunately, accurate solid-precipitation (*P*) measurements have proven nearly impossible to achieve in the Arctic, because precipitation generally falls when it is windy. Gauge undercatch can range from 55–75% depending on the gauge type and wind conditions. The state of knowledge for winter sublimation (*S*) is even more limited. There are few actual measurements and most studies have used physical models to estimate this quantity. Moreover, fundamental questions concerning the boundary-layer physics of arctic winter sublimation remain unanswered. Resolving these is essential to closing local, regional, and pan-Arctic moisture budgets because some studies indicate sublimation may be as much as 50% of the total winter precipitation and 35% of the annual precipitation. This paper summarizes and analyzes the existing literature describing arctic sublimation.
Keywords Precipitation under-catch; snow-drift; snow water equivalents; sublimation

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Introduction

One of the most fundamental components of a surface moisture balance, arctic or otherwise, is *P-E*, the precipitation minus the evaporation. This difference is a measure of the moisture available for runoff and soil recharge, and it indicates, through *E*, the amount of moisture directly recycled to the atmosphere. *P-E* is a fundamental quantity and a key diagnostic of hydrologic systems (e.g. Cullather *et al.* 2000; Serreze and Hurst 2000). Long term changes in *P-E*, produced by shifts in climate, storm tracks or surface vegetation, have profound ramifications for river and stream flow, soil thermal and moisture regimes, vegetation and human utilization of water resources.

P-E appears deceptively simple, easily measured and evaluated, but this could not be further from the truth in the Arctic. There, during the nearly nine months each year when it is winter, temperatures are below freezing, precipitation (*P*) arrives as snow and evaporation (*E*), more correctly called sublimation (*S*) when no liquid is present, returns moisture to the atmosphere directly from snow and ice without it going through the liquid phase (see e.g. Benson and Sturm 1993). The Arctic is also a region of strong and frequent winds (see e.g. Olsson *et al.* 2002). These winds blow and transport snow both during and after snowfall (Sturm *et al.* 2001a). Tundra vegetation covers most of the region, so precipitation gauges cannot be located in sheltered forest clearings as is commonly done in lower latitudes. Exposed to high winds and blowing snow, gauge inaccuracies and failures are the rule and it has proven difficult to develop reliable and accurate arctic precipitation gauge networks.

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A voluminous literature spanning more than five decades attests to this continuing struggle to monitor *P* in the Arctic. Add to this the low human population, a sparse land-based station network and virtually non-existent sea-ice stations, and the result is an extremely limited and generally inaccurate knowledge of precipitation (*P*) throughout the pan-Arctic.

The state of knowledge for winter evaporation/sublimation (*E* or *S*) is even worse. *S* is not routinely monitored anywhere. The few measurements that exist come from research projects, not monitoring programs, and they are spread widely in both space and time. More often than not, these are not measurements at all, but physical model results subject to modeling errors. Moreover, as discussed below, fundamental questions concerning Arctic winter sublimation rates and conditions remain unanswered. Resolving these are essential to close local, regional, and/or pan-Arctic moisture budgets, because the evidence indicates that sublimation (*S*) is a substantial fraction (between 10 and 50%) of the total winter precipitation. In fact, our Alaskan field observations suggest that sublimation can be as large as 100% in some regions. For example, significant snow-free areas exist on winter satellite imagery and we hypothesize they are the result of high sublimation due to katabatic winds (Figure 1). Since winter precipitation supplies up to 80% of the annual total runoff from many Arctic drainage basins (e.g. McNamara *et al.* 1998), the importance of improving the understanding and estimates of *S* is clear.

In the Arctic, *P-S* takes on additional importance because it determines the amount of snow covering the ground or sea ice during the winter. This snow cover is effectively the upper land or ocean surface for nine months of the year, and as a result it controls the surface albedo and affects the energy exchange between land, sea ice, ocean and atmosphere (see e.g. Kane *et al.* 1991; Liston 1995, 1999, 2004; Everett *et al.* 1996; Hinzman *et al.* 1996, 1998; McNamara *et al.* 1997, 1998, 1999; Nelson *et al.* 1998; Neumann and Marsh 1998; Marsh 1999; Persson *et al.* 2002; Uttal *et al.* 2002).

What can be done to improve the knowledge of arctic winter *P-S*? We could improve the instrumentation and protocols used to measure *P* on a systematic basis. Work of that nature is underway and is essential but, based on past research history, any future progress is likely to be slow. An equally useful strategy is to focus on *S*, resolving the contradictions that exist in our current state of knowledge and improving our ability to model sublimation rates from

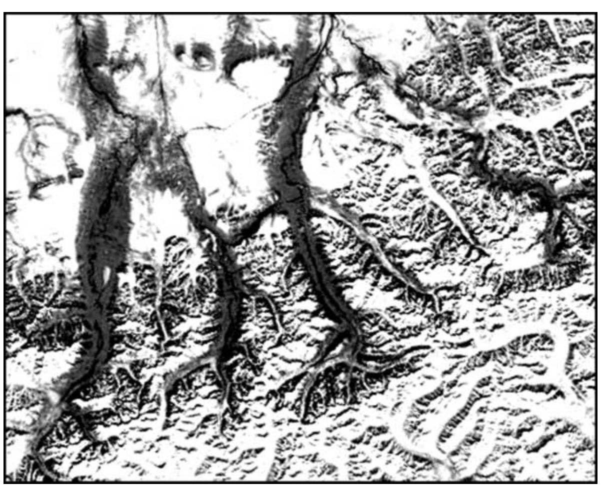

Figure 1 End of winter, Landsat 7 image showing dark, snow-free valleys on the north side of the Brooks Range, in Arctic Alaska. The reduced snow cover in these areas is thought to be the result of strong katabatic winds that flow down the valleys towards the north (top of the figure). The center of this image is at approximately 68°30′ N, 148°50′ W

readily available meteorological data. This represents a promising method of improving our ability to estimate P at the local to the pan-Arctic scale.

Wherever winter temperatures are consistently below freezing (like the Arctic), conceptually there is a direct linkage between snow-water-equivalent depth on the ground (D), sublimation (S), and precipitation (P):

$$D = P - S \tag{1}$$

A number of important assumptions and constraints must be invoked to write this balance in this way, chief of which is that we have assumed local spatial variations in erosion/deposition are negligible, and that we can measure or estimate S. With these assumptions, Equation (1) allows easier-to-measure or model values (D and S) to be used to compute a hard-to-measure quantity (P). At the same time, evaluating P in this manner provides insight into the two other critical winter components of the surface moisture budget.

The winter moisture balance

The struggle to measure Arctic winter precipitation (*P*)

In the Arctic, accurate solid-precipitation measurements have proven difficult to achieve because precipitation generally falls when it is windy. In most cases, winds lead to a significant underestimate of the precipitation (Black 1954; Benson 1982; Larson and Peck 1974; Goodison *et al.* 1981; Yang *et al.* 1998, 2000), though, in exceptionally windy conditions horizontally blowing snow can be caught by the gauges and lead to an overestimate (Yang and Ohata 2001). Even the most advanced and complex shielded precipitation gauges, such as the World Meteorological Organization (WMO) Solid Precipitation Measurement Intercomparison "octagonal vertical Double Fence Intercomparison Reference" (DFIR) is unable to measure the true precipitation (liquid or solid) and requires corrections for wind speed (Yang *et al.* 1998). This fundamental difficulty was recognized more than 60 years ago (Brooks 1938; Black 1954; Wilson 1954) and remains unchanged today.

The sizes of the corrections are not small. Based on a number of gauge inter-comparison projects (Yang *et al.* 2000), the standard 8-inch diameter unshielded National Weather Service (NWS) precipitation gauge used throughout the USA has an undercatch as high as 75% in windy environments (Goodison *et al.* 1981; Benson 1982; Yang *et al.* 1998, 2000). The shielded Wyoming gauge, also used in the USA (Rechard and Larson 1971; Benson 1982; Goodison *et al.* 1998; Yang *et al.* 2000), has an undercatch estimated at 6% for snowfall without wind and 55% for blowing snow.

In addition to these accuracy problems, Arctic precipitation networks are very sparse. Currently, for the nearly 400 000 km^2 comprising Arctic Alaska (land north of 66°), there are seven unshielded or Alter-shielded gauges (R. Thoman, NWS-Fairbanks, personal communication, 2002), and nine Wyoming gauges in service (R. McClure, NRCS-Anchorage, personal communication, 2002). On average, each one of these 16 gauges yields a less-than-accurate value of winter precipitation for every 25 000 km^2 of this Arctic region. Over the more extensive areas covered by sea ice north of Alaska, no routine precipitation measurements are made (ice-mass-balance buoys deployed as part of the International Arctic Buoy Programme (IABP) do include a sonic sounder to measure snow depth (B. Elder, CRREL, personal communication, 2003)). Scientists have pointed out the inadequacy of this network for climate studies and hydrologic assessment (e.g. Vörösmarty *et al.* 2001, 2002). They have also strongly advocated increasing the network density and improving its accuracy, but the reality is that the number of Arctic stations where precipitation is measured is actually declining rather than increasing (Shiklomanov *et al.* 2002). Faced with this reality, an alternate approach to precipitation measurement is essential in order to make progress on understanding the past, present and future characteristics and evolution of the Arctic freshwater cycle.

Snow-water-equivalent depth (*D*) variability

In Arctic Alaska, we have shown that we can measure *D* on the ground (over the last 10 years we have traveled over 6500 km, making over 170 000 snow-depth measurements on land and sea ice (e.g., König and Sturm 1998; Sturm *et al.* 2001b; Taras *et al.* 2002; Sturm and Liston 2003)). We have found that, due to the frequent occurrence of blowing and drifting snow, the snow-depth or water-equivalent distribution can be very heterogeneous (Liston and Sturm 2002; Sturm and Liston 2003). Snow is eroded from higher-wind-speed areas like ridge tops and accumulates in lower-wind-speed areas like the lee of ridges, topographic depressions and taller vegetation (Benson and Sturm 1993; Pomeroy *et al.* 1993; Liston and Sturm 1998; Sturm *et al.* 2001a, b, c; Liston *et al.* 2002). Over smaller distances (<10 m), dunes and sastrugi create additional snow-depth variations (Mellor 1965).

Where the terrain is flat and there are no drift traps, the lateral transport of snow will produce no net change in the local mean *D* because saltation and suspension will bring as much snow into an area as it moves out. Under these conditions, Equation (1) is valid and *P* can be computed directly from a knowledge of *S* and the local mean value of *D*. For conditions where there is non-zero net lateral transport, Equation (1) will include additional transport terms that must be evaluated using a blowing-snow model like SnowTran-3D (Liston and Sturm 1998) or the spatially distributed version of the Prairie Blowing Snow Model (PBSM; Essery and Pomeroy 1999). This complicates the calculations and likely reduces the accuracy of the computed *P* values.

To determine the regional winter precipitation (*P*) patterns over a 20 000 km^2 area of Arctic Alaska, Liston and Sturm (2002) used a methodology based on Equation (1), with the transport terms added. To include conditions both with and without lateral transport, they used the SnowTran-3D blowing-snow model as part of their procedure. SnowTran-3D accounts for the following: (1) the wind-flow forcing field; (2) the wind-shear stress on the surface; (3) the transport of snow by saltation; (4) the transport of snow by turbulent suspension; (5) the sublimation of saltating and suspended snow (the part of *S* in Equation (1) that is not due to static-surface sublimation); and (6) the accumulation and erosion of snow at the snow surface (*D* in Equation (1)). *P* has been determined for other locations using the Liston and Sturm (2002) methodology, including Alaska (Liston and Sturm 1998), Colorado and Wyoming (Greene *et al.* 1999; Hiemstra *et al.* 2002), Norway (Bruland *et al.* 2004), and Greenland (Hasholt *et al.* 2003), suggesting that the method is, operationally at least, suitable for a pan-Arctic domain.

Sublimation (*S*) uncertainties

Quantifying sublimation (*S*) amounts from both static snow surfaces and from suspended snow particles is inherently more difficult than measuring the snow on the ground. As part of his Wyoming research, Tabler (1975) introduced the idea of a snow particle transport distance: the distance an average-sized snow grain can travel before it completely sublimates. Tabler wrote:

> "Ask any Wyomingite what happened to yesterday's 15-cm snowfall, and chances are good that Nebraska will be named as the most likely recipient. Although this may seem a logical deduction, since the snow was last seen heading east with the wind, it seems unlikely considering the complete absence of any acknowledgement from Nebraska. The difficulty of concealing such big piles of snow rules out Wyoming's stashing away the white stuff west of the border; just sweeping up all of a winter's snow from the 30-km treeless flank of Sherman Hill would bury Cheyenne in 10 meters of snow. Certainly, such an accumulation should be visible from quite a distance, even if stored in a less populous area. The unsuccessful search for these mountains of snow leads to the inescapable conclusion that a third party must be

making off with the winter jewels. Many years of sleuthing have finally exposed evaporation as the villain – the same character well known for robbing fields and reservoirs of precious liquid in the summer. In the great snow rip-off the arch villain, alias 'sublimation,' does most of his work at sub-freezing temperatures as the snow particles are blown along by the wind. The effectiveness of this unique modus operandi is evidenced by the theft having gone unnoticed for so many years."

For Wyoming, Tabler (1975) found snow particle transport distances of approximately 3 km. By applying the same ideas to arctic Alaska, Benson (1982) calculated transport distances of 2–3 km. A tacit assumption underlying these calculations is that the highest rates of sublimation occur when snow particles are moving rather than at rest. In his pioneering work, Schmidt (1972) suggested (based primarily on theoretical considerations) that high sublimation rates were achieved during blowing-snow events because of the high snow particle surface-area to mass ratios and the high ventilation rates achieved when the particles were in the wind stream. He calculated that blowing-snow sublimation rates were two orders of magnitude higher than those from a static snow surface under the same atmospheric conditions (Schmidt 1982). Additional work (Lee 1975; Tabler 1975; Schmidt 1982; Pomeroy and Gray 1995) has expanded our knowledge of the dependence of blowing-snow sublimation rates on air temperature, humidity deficit, wind speed and particle size distribution.

Using Schmidt-type sublimation models, investigators (Benson 1982; Liston and Sturm 1998, 2002; Essery and Pomeroy 1999; Pomeroy and Essery 1999; Hirashima *et al.* 2004) have found that 15–50% of Arctic snow cover is returned to the atmosphere by sublimation. For example, Pomeroy and Gray (1995) found that 15–41% of the winter precipitation was returned to the atmosphere on the Canadian prairies, while in the western Canadian Arctic the values ranged from 20–47% (Pomeroy *et al.* 1997; Essery and Pomeroy 1999). All of these estimates are a direct consequence of the vigorous winds transporting snow because Schmidt-type models do not include sublimation from a static snow surface.

In contrast, other studies (e.g. Mann *et al.* 2000; Déry and Yau 2001, 2002; King *et al.* 2001) have suggested that blowing-snow sublimation contributes considerably less to the total winter amount than sublimation from a static snow surface. In these studies, during blowing-snow events the atmospheric boundary layer (ABL) rapidly saturates with moisture due to losses from the grains themselves, and this significantly reduces ABL sublimation. For example, in a recent modeling study, Déry and Yau (2002) found rates of static snow surface sublimation were 5–10 times higher than those due to blowing snow. Their model (PIEKTUK-D) allows for ABL saturation through a thermodynamic sub-model. Their highest rates of sublimation were over the sea ice of the Arctic Ocean, yet even those were only about half of those computed using a Schmidt-type model (based on a comparison of rates from Liston and Sturm (2002) for the Kuparuk Basin with those of Déry and Yau (2002) for the same latitude band (60° to 70°)).

Measured values of sublimation support both approaches. Pomeroy and Essery (1999), using eddy-correlation measurements, found good agreement between their measured sublimation rates during a blowing-snow storm and the results of their blowing-snow model, PBSM, which contains only a blowing-snow sublimation algorithm (Schmidt-type formulation). In contrast, Déry and Yau (2002) also found good agreement between their model results (which emphasize static-surface sublimation) and measurements from the MacKenzie Basin, although this tended to be a low-wind environment. The models could be tested using other measurements, including data from Antarctica (e.g. King *et al.* 1996, 2001; Bintanja 1998; Mann *et al.* 2000), but these data are scattered throughout a wide literature, or in many cases, are unpublished. Moreover, the available Arctic and Antarctic data sets vary in quality, and have never been collected, evaluated and used in an all-encompassing, systematic fashion to address key sublimation questions.

It is possible that no contradiction exists between the two types of sublimation: There may be two distinctive sublimation regimes, with the nature of the ABL interaction with the overlying air mass determining which regime dominates. When the ABL is unable to mix vertically, or mixes with air that is already saturated, then blowing-snow sublimation is likely to be minimal. In this "humid" regime, surface sublimation would also be limited, unless the general relative humidity of the ABL tended to drop between the blowing-snow events. Arctic sea ice is one place where this regime is likely to exist. Andreas *et al.* (2002) reported that, during the Surface Heat Budget of the Arctic Ocean (SHEBA) project (Perovich *et al.* 1999), the ABL was nearly always saturated (with respect to ice) due to the presence of leads and open water. Observations made at SHEBA during about the same time (Sturm *et al.* 2002) suggested that water vapor from the air was being deposited on the snow surface covering the sea ice rather than sublimating away. In contrast, a "dry" regime exists when the ABL is able to mix continuously with an overlying pool of dry air, removing moisture from the near-surface layer of air and snow particles. In this case, blowing-snow sublimation undoubtedly exceeds static-surface sublimation for all the reasons put forward by Schmidt (1972) and simulated by the Schmidt-type models. This "dry" regime is supported by our Arctic Alaska winter fieldwork, where the vast majority of days with blowing snow occurred under blue-sky conditions.

This two-regime model of sublimation is still hypothetical and will require careful measurements for confirmation. Even if this model proves to be correct, there will still be questions that need to be answered in order to estimate S reliably. For example, how do we determine which regime exists at a given location or under a given set of conditions?

Other studies of static surface sublimation over the Greenland ice sheet have calculated total annual ice sheet sublimation to be 12% or 23% of the precipitation, depending on the method used (Box and Steffen 2001). Ohmura *et al.* (1999) estimated that evaporation/sublimation effectively reduced the annual ice sheet precipitation input by 12.5%.

There is still another complication that must be dealt with before reliable Arctic-wide S estimates can be obtained. Trees cover over 30% of pan-Arctic land mass when the region is defined by the watersheds of the large, northward-flowing rivers (R. Lammers, University of New Hampshire, personal communication, 2001). Pomeroy and Schmidt (1993) showed that the boreal forest canopy intercepts up to 60% of winter snowfall. They also note that this snow, exposed more fully to wind and sun than snow beneath the canopy, has annual sublimation losses between 30–40% of annual snowfall, while snow beneath the canopy has very little sublimation. These rates are as high as those for blowing snow, but occur from relatively static snow surfaces. Schmidt (1991) presented a model to account for these sublimation losses that is closely related to his blowing-snow sublimation formulation, and Hedstrom and Pomeroy (1998) presented measurements of these losses from the Canadian boreal forests that confirm the model. Lundberg and Halldin (2001) provide a through review of canopy snow evaporation/sublimation measurements, processes and models. Unfortunately, questions remain as to how sublimation rates vary with forest density, canopy type and atmospheric conditions. Moreover, in the boreal forest areas of the pan-Arctic, appropriate canopy and sub-canopy models must be implemented to correctly determine sublimation, and this means that models must include accurate representations of forest characteristics.

Conclusions

The preceding discussion suggests that significant deficiencies exist in our knowledge of fundamental Arctic winter moisture budget components. In its simplest terms, the winter surface moisture budget (in the absence of any net horizontal transport) is defined by:

snow-water-equivalent depth on the ground (D) equals precipitation (P) minus sublimation (S). These three terms, D, P and S, are the most fundamental components of the winter Arctic hydrologic cycle, and understanding them is essential to understanding arctic moisture-related processes. A primary deficiency is the fact that accurate arctic precipitation (P) measurements have proven difficult to achieve; this has been true in the distant past and will likely continue well into the future. In contrast, snow-water-equivalent on the ground (D) is relatively easy to accurately measure. If S could be measured or estimated, P could be computed. However, we have found that Arctic winter sublimation (S) processes vary widely in both time and space. Sublimation is a function of air temperature, humidity and wind speed variations associated with changing weather patterns, and space-dependent variations related to local surface roughness, vegetation, proximity to open water/ocean and other environmental factors. All of these affect the ABL, altering whether it is saturated or not, and apparently controlling whether sublimation rates peak during or between blowing-snow events.

Thus, a key goal of the Arctic research community should be to investigate the temporal and spatial variability of sublimation. This can be accomplished by measuring sublimation rates using eddy correlation systems, monitoring ABL conditions and describing local vegetation, topography and other surface characteristics over a wide variety of Arctic environments and a wide range of atmospheric conditions. Once the fundamental system components are understood, tools should be developed, using appropriate physically based models, vegetation maps, DEMs and weather records, to allow the simulation of local sublimation (S) variations throughout the Arctic. These S values, in conjunction with snow-water-equivalent on the ground (D) measurements, can then be used to produce alternative precipitation (P) estimates.

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