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Review Article

Difficulties in Determining Snowpack Sublimation in Complex Terrain at the Macroscale

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In many mountainous regions, snowmelt is an essential component of water resources and ecosystem function and snow sublimation often leads to water loss from the given drainage basin. Previous investigators have developed numerous modeling and measurement techniques to quantify sublimation, illustrating high variability over short distances. The complexities of modeling and measuring sublimation limit investigations to smaller scales in complex terrain and therefore the effects that microscale controls on sublimation have at the macroscale are not well understood. A key component of microscale variability, vegetation cover, can change on short time scales relative to other components (e.g., slope, aspect, and elevation) in response to natural and anthropogenic influences such as land use practice, drought, wildfire, insect infestation, and climate change. Basic vegetation-sublimation relationships may vary within a given drainage basin, by climate type, seasonally, and interannually. It is therefore particularly important to advance understanding of vegetation effects on sublimation at the macroscale.

1. Introduction

In mountainous regions with seasonal snow cover, melting snow is generally the major component of water resources, providing runoff during the late spring and early summer when demands from humans and the local environment are increasing [1–4]. Snow accumulation also affects ecosystem processes with impacts on soil moisture, soil porosity, biogeochemical processes, and wildfire [5]. Snow surface melting can only occur if the surface temperature is equal to the melting point with additional energy required for melting, while sublimation occurs when the vapor pressure near the ice crystal surface is greater than that of the environment [6, 7]. When sublimation occurs, there is a direct loss of water from the drainage basin upon advection out of the basin.

Sublimation can be separated into three categories: (1) canopy sublimation, (2) sublimation from ground snow cover, and (3) sublimation from blowing snow [8]. Canopy sublimation occurs when snowfall intercepted by vegetation sublimates before being unloaded to the ground snowpack [9]. Relatively high above canopy wind speeds along with a high surface area to mass ratio of snow retained in the canopy lead to increased sublimation relative to snow stored on the ground [8, 10]. Compared to ground snow cover in open areas, the subcanopy environment typically experiences reduced shortwave radiation and wind speeds as well as increased humidity and longwave radiation; therefore, sublimation from ground snow cover is affected by the presence of forest cover [8]. These components of the surface energy budget can be further altered by the presence of snow retained in the forest canopy [9]. Sun and wind crusts reduce sublimation [11] and the canopy modified surface energy budget also influences the surface snow condition and thus sublimation. Last, blowing snow sublimation occurs when saltating or suspended snow particles are surrounded by air subsaturated with respect to ice, which can occur at very rapid rates [8].

In mountainous terrain, there is high microscale variability in vegetation type, density, forest clearing size, and canopy geometry, as well as elevation, slope, aspect, and curvature [12]. Even site specific measurements of sublimation require careful attention to reduce error [13] and detailed observations of wind, temperature, and humidity in the least (which vary greatly in complex terrain) are required to model sublimation [14]. High resolution fields from interpolation

and modeling have been used to produce spatially distributed simulations of sublimation [8, 15–17] yet many such investigations are only hundreds of square kilometers in scale, and large scale studies are restricted to high latitudes (e.g., [6, 18]).

The goal of this review is to highlight the difficulties that arise when studying sublimation at the macroscale in complex terrain and emphasize important advances that can be made in the knowledge of macroscale sublimation. An overview of common techniques used to measure and model snow sublimation is given followed by a summary of results from previous investigations. Impediments to determining macroscale sublimation in complex terrain are then discussed as well as future directions for the study of macroscale sublimation from complex terrain.

2. Measuring and Modeling Sublimation

In this section, overviews of sublimation measurement and modeling techniques are given. More extensive discussions are given by Lundberg and Halldin [13] for canopy sublimation, Mahrt and Vickers [19] and Hood et al. [14] for ground sublimation, and Groot Zwaaftink et al. [16] for blowing snow sublimation. Therefore, this section is not meant to be an exhaustive review of techniques; rather, various methods are briefly introduced to illustrate the difficulty in their use for determining sublimation at the macroscale in Section 4. In this review, the term sublimation, unless otherwise specified, refers to a mass loss from the snowpack through latent heat flux over a given time period, with the understanding that deposition, condensation, and evaporation can also occur over the given time period.

2.1. Measurement Techniques. First, it is important to note that there are no measurement techniques to separate blowing snow sublimation from other processes in uncontrolled environments, and therefore blowing snow sublimation in complex terrain is only quantified by simulation [16, 20]. Latent heat flux from a snow covered surface is most often measured with eddy-covariance systems (e.g., [9, 21, 22]) or gravimetrically (e.g., [11, 23–26]). For both methods, latent heat flux is not separated by sublimation, deposition, condensation, evaporation, or transpiration and assumptions must be made about the dominant form of latent heat flux based on season [13, 27]. High frequency observations with a sonic anemometer, an instrument that is not ideal for long-term observations in harsh alpine environments, are required for eddy-covariance measurements [14]. To measure sublimation over long periods that experience precipitation, gravimetric techniques require accurate measurements of precipitation, a difficulty due to undercatch [13]. Snow removal by wind and sampling a snow surface with properties that are not representative of the surrounding environment are additional sources of error with gravimetric techniques [11, 13].

Measurements of ion chemistry and water isotopes can be used to quantify sublimation [28]. Cumulative snowfall solute concentrations and isotopic values can be determined upon regularly sampling snowfall events and measurements from the snowpack can then be used to quantify the depth of sublimation provided there is no significant mass loss from melting [28]. In a small catchment in Idaho, USA, isotopic values were measured from streamflow by Koeniger et al. [29] suggesting that variation in isotopic enrichment was due to variation in sublimation due to forest cover characteristics.

2.2. Modeling Techniques. There are numerous approaches to modeling sublimation that vary in complexity based on the level of measurement detail for input and the sublimation type of interest (i.e., ground, canopy, or blowing snow sublimation). For ground snow sublimation, the most common method is the bulk aerodynamic method, requiring meteorological observations at only one height. The aerodynamic profile method has the potential to be more accurate than the bulk method but requires observations at multiple heights [14, 30]. As an example of a bulk method, the Alpine MUltiscale Numerical Distributed Simulation Engine (AMUNDSEN) employs the following that is valid for stability, roughness, and wind speeds common over snow surfaces in mountainous terrain [8]:

$$\mathrm{LH} = 32.82\,(0.18 + 0.098U)\,(e - e_s), \tag{1}$$

where LH is the latent heat flux in $\mathrm{W\,m^{-2}}$ (positive downward), e_s is the saturation vapor pressure with respect to ice (hPa), e is the vapor pressure (hPa), and U is the wind speed ($\mathrm{m\,s^{-1}}$) at typical measurement levels. The snow surface temperature is required to determine e_s and iterative procedures can be used to determine the surface temperature required for energy balance when the measured air temperature is below 273.16 K [8]. The energy budget equation for the snow surface in a simple one-layer model is given by [31]

$$Q + H + \mathrm{LH} + A + B + \mathrm{LM} = 0, \tag{2}$$

where Q is net radiation, H is the sensible heat flux, A is the energy flux from precipitation, B is the soil heat flux, and LM is the energy available for melting. When the measured air temperature ≥273.16 K, the snow surface temperature is assumed to be 273.16 K (e.g., [8]).

Canopy and blowing snow sublimation models fundamentally differ from ground sublimation models to account for the higher surface area to mass ratio of snow stored in the canopy and suspended snow particles [10, 16]. Sublimation from an ice-sphere in thermodynamic equilibrium with the environment [32, 33] is given by [10, 34]

$$\frac{dm}{dt} = \frac{2\pi r\,(\mathrm{RH}/100 - 1) - S_p\Omega}{h_s\Omega + 1/D\rho_v\mathrm{Sh}} \tag{3}$$

with

$$\Omega = \frac{1}{\lambda_t T_a \mathrm{Nu}}\left(\frac{h_s M}{RT_a} - 1\right), \tag{4}$$

where m is the ice-sphere mass (kg), r is the radius (m), RH is the relative humidity (%) with respect to ice, S_p is absorbed solar radiation ($\mathrm{W\,m^{-2}}$), h_s is the latent heat of sublimation ($\mathrm{J\,kg^{-1}}$), D is the diffusivity of water vapor ($\mathrm{m^2\,s^{-1}}$), ρ_v is the saturation vapor density with respect to ice ($\mathrm{kg\,m^{-3}}$),

Table 1: Slightly modified from Jackson and Prowse [26], selected studies of sublimation from snow covered terrain.

Study	Methods	Site type	Sublimation/evaporation
Bernier and Swanson, 1993 [35]	BA, Gr	Forest and open	0.25–1.07 mm dy^{-1}
Doty and Johnston, 1969 [36]	Gr	Open	0.15 (January)–1.56 (April) mm dy^{-1}
Fassnacht, 2004 [37]	BA	Various USA sites	0.23–0.67 mm dy^{-1}
Golding, 1978 [38]	BA	Subalpine forest	1.2 mm dy^{-1} (1975), 2.0 mm dy^{-1} (1976)
Hood et al., 1999 [14]	AP	Alpine	0.9–1.8 mm dy^{-1} (annual 15% max SWE)
Lundberg et al., 1998 [39]	Gr	Forest (Scotland)	Max 3.9 mm in 7 hours (canopy)
Kaitera and Teräsvirta, 1972 [40]	BA	Boreal forest	0.35 (subcanopy)–0.45 (open) mm dy^{-1}
Kaser, 1982 [41]	Gr	Alpine	Mean 0.25 mm dy^{-1}, max 2.0 mm dy^{-1}
Marks and Dozier, 1992 [42]	BA	Alpine	Mean 2 mm dy^{-1}
Martinelli, 1960 [43]	Gr	Alpine	±0.67 mm dy^{-1}
Meiman and Grant, 1974 [44]		Alpine, forest, and open	45–60% snow season precipitation
Molotch et al., 2007 [9]	EC	Subalpine forest	0.41 (subcanopy)–0.71 (canopy) mm dy^{-1}
Montesi et al., 2004 [7]	Gr	Subalpine forest	Canopy, 20–30% of total snowfall
Nakai et al., 1999 [45]	EC	Boreal forest canopy	1.2 mm dy^{-1} (snow covered)
Pomeroy et al., 1998 [10]	BA	Boreal forest	0.41–1.88 mm dy^{-1}
Pomeroy and Essery, 1999 [21]	EC	Prairie	1.8 mm dy^{-1}
Rylov, 1969 [46]	Gr	Open (semidesert)	0.08 (January)–0.6 (April) mm dy^{-1}
Schmidt and Troendle, 1992 [47]		Canadian forest	Annual 46 mm (canopy)
Schmidt et al., 1998 [11]	BA, Gr	Subalpine forest	0.61 (southerly), 0.43 (northerly) mm dy^{-1}
Storck et al., 2002 [48]	Gr, SC	Subalpine forest	100 mm/winter, <1 mm dy^{-1}
Suzuki et al., 2006 [49]	BA, model	Taiga, larch forest	1.0 (forest)–2.0 (open) mm dy^{-1}
West, 1962 [23]	Gr	Forest, subcanopy	50 mm (2.7% of snowfall) annually
Zhang et al., 2003 [50]	BA, Gr	Taiga	0.2–1.0 mm dy^{-1}
Zhang et al., 2004 [25]	BA, Gr	Taiga, larch forest	0.22–0.32 mm dy^{-1}

AP: aerodynamic profile; BA: bulk aerodynamic; EC: eddy covariance; Gr: gravimetric, lysimeter; SC: snow course.

Sh and Nu are the Sherwood and Nusselt numbers (nondimensional), respectively, which depend on the wind speed and particle size, λ_t is thermal conductivity of the air (J m^{-1} s^{-1} K^{-1}), T_a is the air temperature (K), R is the universal gas constant (8313 J kmole^{-1} K^{-1}), and M is the molecular weight of water (18.01 kg kmole^{-1}). Complete formulae and definitions for all terms in (3) and (4) are given by Pomeroy et al. [10] and Liston and Elder [34]. To get a canopy sublimation rate (mm s^{-1}), (3) can be multiplied by m^{-1} (kg^{-1}); the intercepted snow load (kg m^{-2}) that is dependent on leaf area index; and a nondimensional coefficient to account for the exposed snow surface area being less than that of the individual snow grains, dependent on snow load and leaf area index [7, 34, 51].

In modeling blowing snow sublimation, an ice-sphere sublimation model (e.g., (3)) is commonly used to determine mass loss from individual particles in a size spectrum of suspended particles [16, 52, 53]. Three-dimensional models are ideal for simulating blowing snow in complex terrain and include SnowTran-3D [34], SYTRON3 [54], and Alpine3D [55]. Of these three commonly used three-dimensional snow transport models, only Alpine3D explicitly implements negative feedback processes involving sublimation (i.e., in a plume of snow particles, sublimation reduces temperature and snow mass concentration and increases humidity [16]).

Valeo et al. [56] developed a method to estimate sublimation by use of a Global Positioning System. Using a relationship between zenith wet delay and specific humidity, a bulk aerodynamic method can be used to estimate sublimation by replacing measurement height humidity observations with Global Positioning System derived precipitable water [56]. Good agreement between observations from sublimation pans and modeled sublimation has been reported with this method [56].

3. Previous Research Efforts

3.1. Ground and Canopy Sublimation. Jackson and Prowse [26] present a table of sublimation estimates derived from 30 previous studies around the world indicating that sublimation varies considerably both spatially and temporally (Table 1). As seen in Table 1, sublimation rates are generally highest in wind exposed high altitude environments (e.g., a mean of 2 mm dy^{-1} from Marks and Dozier [42] and 0.9–1.8 mm dy^{-1} from Hood et al. [14]). Ground sublimation from open areas is generally more rapid than adjacent forested areas (e.g., from Table 1, 1.0 versus 2.0 mm dy^{-1} from Suzuki et al. [49] and 0.35 versus 0.45 mm dy^{-1} from Kaitera and Teräsvirta [40]). In mountainous terrain in southwest Idaho, Reba et al. [22] measured sublimation from ground

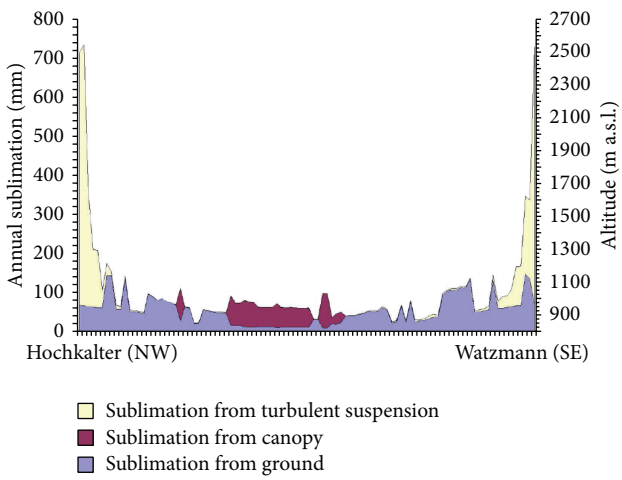

Figure 1: Reproduced from Strasser et al. [8], simulated contributions to total annual snow sublimation from ground sublimation, canopy sublimation, and blowing snow sublimation along a ~4 km transect across Wimbach valley in Berchtesgaden National Park, Germany.

snow cover during three winter seasons at a wind sheltered site (aspen forest) and wind exposed site (nonforested). Sublimation was less than 10% of maximum snow water equivalent (SWE) for all three winters at the sheltered site and ranged from 16% to 41% of maximum SWE at the exposed site (with blowing snow sublimation potentially contributing to the high sublimation at the exposed site).

Canopy sublimation rates in boreal forest are considerably larger than ground sublimation rates, from 1.2 mm dy^{-1} by Nakai et al. [45] for canopy sublimation to 0.45 mm dy^{-1} in forest clearings by Kaitera and Teräsvirta [40]. In a subalpine Colorado forest, Molotch et al. [9] measured canopy sublimation rates of 0.71 mm dy^{-1} and subcanopy sublimation rates of 0.41 mm dy^{-1} during March and April. Jackson and Prowse [26] found sublimation rates of 0.4 mm dy^{-1} at open sites during the snowmelt period in a small catchment in the Okanagan River basin in southern British Columbia. On adjacent north and south facing forested slopes in central Colorado, Schmidt et al. [11] found that ground snow cover sublimation is 20% of maximum SWE, with 1.2 times more sublimation on the south facing slope.

In a modeling study, Strasser et al. [8] produced a figure of sublimation totals along a mountain-valley transect in Berchtesgaden National Park, Germany (Figure 1). It can be seen that canopy sublimation is generally the dominant form of sublimation in tree-covered areas although there are portions of the transect with lower leaf area index (not shown) where ground sublimation is larger than canopy sublimation. Strasser et al. [8] report total vapor loss to be as low as 10% of annual snowfall in relatively open canopies with low wind speeds. In addition to leaf area index, the duration of snow cover held within tree canopies greatly influences the prominence of canopy sublimation across a landscape. Yamazaki et al. [57] highlight that solid water storage in the canopy is continuous in Siberian taiga forest from October through March and note that this contrasts with temperate regions where snow storage in the forest canopy is intermittent except in areas with frequent snowfall.

Cold temperatures can limit sublimation which is evident in Table 1 with low sublimation rates during winter months. This is particularly evident in the values reported by Rylov [46] for an open semidesert area in Kazakhstan with January sublimation rates of 0.08 mm dy^{-1} compared to 0.6 mm dy^{-1} in April. Regions that experience intensely cold winters tend to have little sublimation during the winter [57, 58]. Yamazaki et al. [57] report modeled canopy sublimation rates of <0.05 mm dy^{-1} during midwinter in Siberian taiga forest. Zhou et al. [58] report ground sublimation of only ~5 mm in the Gurbantunggut Desert, China, over ~100-day period. These low sublimation rates in cold regions should not be considered negligible as low winter precipitation results in 23.63% of snowfall sublimated in the Gurbantunggut Desert [58]. This is comparable to relatively high energy subalpine areas of the southwest USA where 17–30% of annual snowfall is lost to sublimation [28]. Of total annual arctic precipitation, 35–50% is thought to be lost to sublimation from tundra [59].

Previous investigators have also noted mass increase through condensation/deposition exceeding sublimation/evaporation, often during warm periods and through entire snowmelt seasons (e.g., [14, 24, 26]). Rates of mass increase as high as 0.75 mm dy^{-1} for March 2007 were measured at a forested site in southern British Columbia by Jackson and Prowse [26]. Subcanopy adjustments to humidity measurements taken at open sites often include setting the relative humidity to 100 percent when the snowpack is melting (e.g., [31]). Strasser et al. [31] simulate nearly equal magnitudes of sublimation and deposition for dense forest cover in Berchtesgaden National Park, Germany, for a single entire snow season (2005/2006). Hood et al. [14] report a May–July mass increase through latent heat flux in 1995 with an increase of 17 mm in May alone at an alpine site in Colorado. Strasser et al. [8] simulated deposition greater than 200 mm in a single year where snow cover persisted into summer in Berchtesgaden National Park, Germany.

Several investigators have highlighted the importance of a few high sublimation events to total seasonal sublimation (e.g., [14, 60]). Sublimation estimates from the aerodynamic profile method in the Front Range of the Rocky Mountains in Colorado, USA, by Hood et al. [14] indicate a two-day period with sublimation that totaled 21% of the monthly total. High sublimation events occur with high wind speeds along with relatively warm temperatures and low specific humidity leading to large vapor pressure deficits (Figure 2). While sublimation is often episodic, there are consistent diurnal patterns as well (Figure 3) with positive net radiation and higher snow surface temperatures, leading to sublimation during the day, and negative net radiation and low snow surface temperatures resulting in latent heat flux to the snowpack (deposition/condensation) at night.

3.2. Blowing Snow Sublimation. Blowing snow sublimation varies greatly in space and time, partly because high wind speeds are required to transport significant amounts of snow

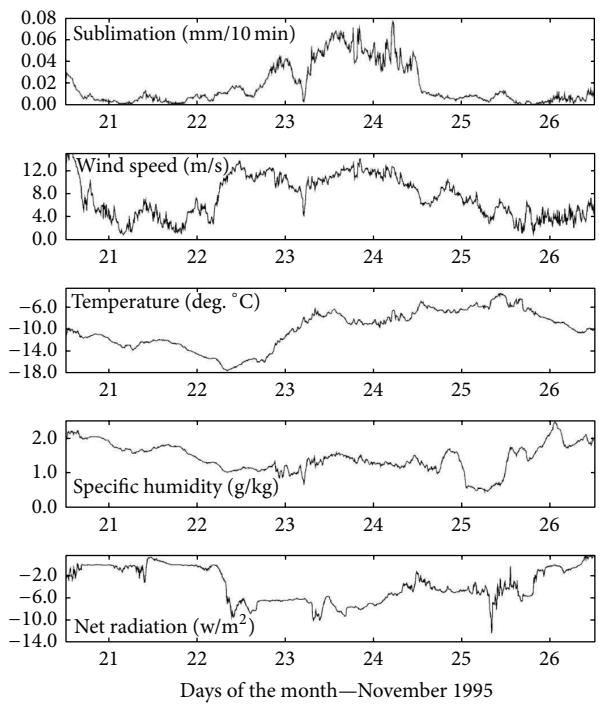

Figure 2: Reproduced from Hood et al. [14], ten-minute averages of sublimation, wind speed, temperature, specific humidity, and net radiation from Niwot Ridge, Colorado, USA. Tick marks denote the beginning of the day.

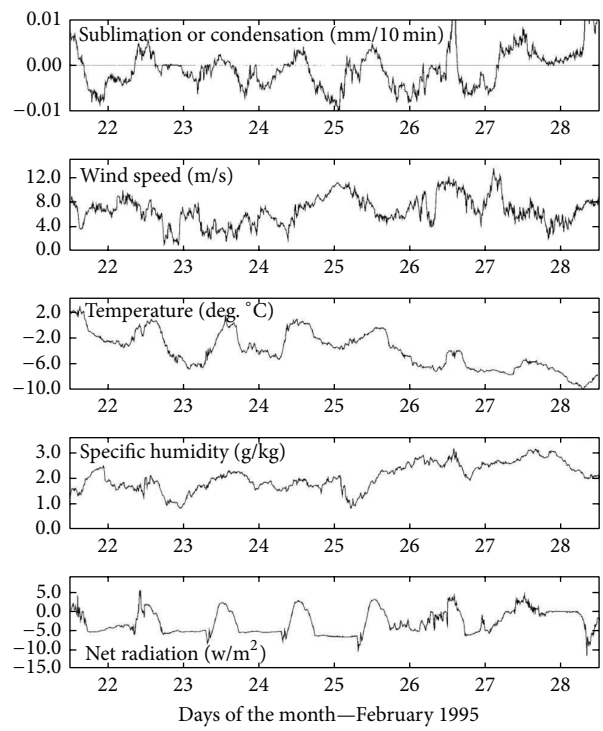

Figure 3: Reproduced from Hood et al. [14] as in Figure 2 but for seven days in February. Condensation is represented by negative values in the top panel.

[20, 61]. In a rugged alpine and subalpine region, Strasser et al. [8] show that seasonal totals of blowing snow sublimation can greatly exceed canopy and ground sublimation for exposed alpine ridges (Figure 1) representing a vapor loss that is 90% of total snowfall, but only accounting for a 4.1% loss of snowfall over the entire 210 km^2 study area in Berchtesgaden National Park, Germany. Bernhardt et al. [17] expanded on the Strasser et al. [8] study by including a gravitational snow transport scheme that resulted in more snow transport from wind exposed areas (e.g., crests and steep terrain) to lower wind sheltered areas. This reduced sublimation, as a percent of total snowfall, to 1.6% [17]. Groot Zwaaftink et al. [20] find even smaller contributions (0.1% of winter snowfall) of blowing snow sublimation to the snow water budget over a 2.4 km^2 alpine area in the Swiss Alps. The Alpine3D model used by Groot Zwaaftink et al. [20] includes feedback processes that make blowing snow sublimation self-limiting which is one reason why estimates are lower than those of Strasser et al. [8] and Bernhardt et al. [17]. Groot Zwaaftink et al. [20] also find that significant blowing snow sublimation is highly episodic, restricted to short time intervals that often correspond to Föhn storms, and that blowing snow sublimation can increase by an order of magnitude over localized areas of the alpine tundra. In a drier climate than the Swiss Alps, the Canadian Rockies, MacDonald et al. [62] simulated blowing snow sublimation to be between 17 and 19% of seasonal snowfall along a largely treeless ridge. In a relatively windy 65 km^2 catchment of the low arctic (e.g., a monthly mean 10-m speed of 7 $m\,s^{-1}$) where fetch is longer than most alpine areas, Pomeroy et al. [63] simulated blowing snow sublimation to be 19.5% of annual snowfall.

3.3. Sensitivity to Changes in Forest Cover. While most elements of microscale variability remain constant on short time scales (i.e., slope, aspect, curvature, and elevation) vegetation patterns can vary considerably through natural (wildfire, bark beetle infestation) and anthropogenic (land management practices) processes [5, 64]. Therefore, understanding how changes at the stand scale impact snow sublimation at the macroscale is critical from a future water resources perspective. LaMalfa and Ryle [5] note declining aspen (*Populus tremuloides*) stands in the western USA attributable to fire suppression, ungulate browsing, land use practices, and climate change. LaMalfa and Ryle [5] observed greater water yield for melt driven runoff and groundwater recharge in aspen stands compared to mixed conifer stands and noted the desire for Western USA land managers to restore aspen forests suggesting that the economic benefits from increased water yield could eventually overcome restoration costs.

It is likely that vegetation impacts on sublimation vary by region. Pugh and Small [64] examined sublimation differences among living and dead (from bark beetle infestation) lodgepole pine (*Pinus contorta*) stands in Colorado, USA, concluding more total sublimation loss from live stands. Pugh and Small [64] admit that these results may differ in regions with storm characteristics more conducive to canopy interception and interstorm characteristics less conducive to rapid

sublimation. Similarly, it is possible that the net effect of dense forest cover on sublimation in cold regions is opposite to that in warmer forests due to differences in the length of time during which snow is held in the canopy [57] and differences in the importance of canopy sheltering on the snowpack energy budget. For example, Harpold et al. [65] observed reduced interception in recently burned forest compared to unburned forest in the Jemez Mountains of New Mexico, USA, yet peak snow accumulation was observed to be greater in unburned areas. Harpold et al. [65] speculated that, in a region with high solar elevation angles and frequent cloud-free conditions, ground sublimation may be especially sensitive to increased shortwave radiation following the removal of the forest canopy by fire. Furthermore, if short canopy snow cover duration is a limiting factor for seasonal canopy sublimation in warmer regions, then changes in forest radiation and wind sheltering impacts on ground sublimation may overcome reduced intercepted sublimation upon forest thinning.

4. Challenges in Determining Macroscale Snow Sublimation

The few studies of sublimation across large scales are restricted to high latitude regions. Essery et al. [18] examined canopy sublimation from the Northern Hemisphere boreal forests through coupling a GCM with a canopy snow process model. The fraction of sublimated snow for grid cells was most commonly just greater than one-fifth, with a few grid cells in the continental climates of Asia displaying fractions of one-half [18]. Déry and Yau [6] used the European Centre for Medium Range Weather Forecasts Reanalysis to determine ground snow sublimation through a bulk aerodynamic method and blowing snow sublimation through a double-moment blowing snow model for Antarctica and the Arctic. They found that ground snow sublimation is 7% of annual precipitation over the entire Mackenzie River Basin in Canada and more than 17% of annual precipitation over Antarctica. Van den Broeke [66] used a general circulation model and a bulk aerodynamic method to determine that sublimation over Antarctica represents 10–15% of the annual precipitation.

In mountainous regions of the midlatitudes, terrain is highly complex, with elevation, slope, aspect, curvature, vegetation type, forest density, forest clearing size, and canopy geometry all varying greatly over small spatial scales [12]. It is therefore difficult to scale up results of measured sublimation from specific sites to the basin scale [26]. Similarly, modeling sublimation in complex terrain must be done at high resolution and therefore multiple meteorological variables must be represented at high resolution to do so (e.g., [67]). Interpolated meteorological fields from observations [8, 15, 16, 67] and high resolution fields from mesoscale models [17] can be used to produce spatially distributed simulations of sublimation. Unfortunately, due to the requirement for high resolution modeling, the many investigations concerned with spatially distributed simulations of sublimation in complex terrain are small in scale, only hundreds of square kilometers (e.g., [8, 15–17, 20]).

Isotopic values and solute concentrations of streamflow can provide valuable information on sublimation (e.g., [29]) but seasonal quantification of sublimation at the macroscale would require detailed observations at high resolution as Ohlanders et al. [68] note that snowpack isotopic values vary considerably by elevation and at the microscale with constant elevation. Temperature influences the isotopic variability of precipitation, requiring that samples of precipitation composition be taken in multiple areas of a macroscale study area. Last, isotopic compositions also change during transport as surface and subsurface flow [68], making the contribution from sublimation difficult to isolate.

From a water budget prospective, total macroscale sublimation over time period t can potentially be estimated by considering the following water budget equation:

$$F = P - \mathrm{SUB} - E - T - \Delta\mathrm{ST}, \tag{5}$$

where F is measured river discharge over time period t, P is total precipitation over the contributing basin area, SUB is the total sublimation over the area, E is evaporation, T is transpiration, and ΔST is the increase in water storage over the period t. If t is a long time period (e.g., multiple years), then ΔST will be considerably less than all other terms and errors related to runoff lag will be minimal. Even with large t, errors in sublimation determined from (1) are sensitive to errors in macroscale estimates of P, E, and T in addition to errors in measured river discharge. For short time periods (e.g., t equals several months), ΔST must be determined and errors related to the transport time of water through the basin become significant [69].

5. Future Directions

Strasser et al. [31] simulate sublimation from an idealized mountain with regular patterns of forested and nonforested areas for three consecutive winters. Among many important findings, Strasser et al. [31] concluded that (1) the effects of aspect on snow cover in forests are only apparent for lower values of leaf area index and (2) sublimation losses from the canopy are relatively more dominant during the dry winter, in contrast to the wet winter when more snow accumulated subcanopy. These results from the Strasser et al. [31] experiment highlight important components of spatial and interannual variability in the forest canopy impact on snowpack. Without accurate measurements of sublimation at the macroscale, it is not possible to validate macroscale estimates of sublimation through simulation. Although simulated magnitudes of sublimation at the macroscale cannot presently be taken as truth, advances in the understanding of how microscale variability affects total macroscale sublimation can still be achieved through careful model experiments (e.g., [31]), especially if the sensitivity of results is examined through parameter adjustment and multimodel simulations. For example, examining model output across forest density gradients controlling for topographic characteristics and/or incrementally changing the canopy characteristics between simulations can quantify the impact of forest cover changes (e.g., thinning, insect infestation, and tree species change).

High resolution (e.g., 30-m) terrain and forest cover data and climate reanalysis datasets that provide the necessary variables to determine latent heat flux from (1) to (4) can be used to estimate sublimation at large scales in complex terrain. For example, in the United States, the National Land Cover Database 2011 [70], National Elevation Dataset [71], and the North American Regional Reanalysis by Mesinger et al. [72] can be used in conjunction with established terrain and vegetation adjustments to spatially distribute meteorological observations based on slope, aspect, elevation, terrain curvature, and forest cover (e.g., [31, 34, 67, 73]). With a large study domain and limited computational resources, the terrain and forest cover data can be coarsened, although important features of topography and forest cover (e.g., steep slopes, dense forest cover adjacent to open areas) may be neglected from coarsening. This can be overcome by defining discrete classes, or by using clustering algorithms, to group high resolution cells based on slope, aspect, elevation, terrain curvature, and forest cover characteristics. Snow model systems can be run once for each group, and the results spatially distributed across the domain (e.g., [74]).

To lend confidence to conclusions drawn from seasonal simulations of sublimation at the macroscale, the spatially distributed meteorological observations and modelled sublimation can be compared to measured values reported by the numerous site specific investigations. Multiple years of meteorological observations and eddy-covariance measured latent heat flux over open areas and subcanopy are reported by Reba et al. [22] and Molotch et al. [9] report subcanopy and above canopy eddy-covariance measured latent heat flux for a single spring. Eddy-covariance flux tower reports of latent heat flux measured above canopy and local meteorological observation are presented particularly well by Broxton et al. [27] for more than four years at two sites in the Rocky Mountains of the USA.

Increased confidence in macroscale estimates of snow sublimation through simulation can also be gained by comparisons between modelled and observed macroscale SWE and snow cover extent. Snow cover duration agreement is most critical but when comparisons between sublimation and melt are of interest, peak SWE agreement is also necessary. The seasonal evolution of basin scale SWE and snow cover extent can be quantified from satellite observations (e.g., MODIS on NASA's Terra and Aqua satellites for snow extent and AMSR-E on the Aqua satellite for SWE). Daily true color imagery and the climate-modeling grid-level-3 data (MOD10C1) at 0.05° resolution based on 500 m Terra MODIS observations [75] can be used to control for cloud cover and to quantify snow covered area. The daily resolution of MODIS derived products makes many observations of snow covered area soon after snow events (given cloud-free conditions) and through spring ablation seasons feasible. In complex forested terrain, however, both SWE and snow covered area tend to be underestimated by products derived from satellite observation (e.g., [76–78]). Snow cover may be completely obscured in areas that are more than ~60% forested and simple adjustments to fractional snow covered area based on canopy cover fraction [78] assume that the canopy obscured snow cover fraction is equal to the visible snow cover fraction. Therefore, automated (e.g., SNOTEL in the USA) or manual (e.g., NRCS snow course or NWS COOP measurements in the USA) ground based measurements should be used to supplement remotely sensed observations. Such observations are often limited in spatial extent and SNOTEL sites are often located in forest openings and in areas favored for snow accumulation [79]. Therefore, when possible, in studies determining macroscale sublimation through simulation, both ground based and remotely sensed snowpack observations should be used to validate the simulated evolution of SWE and snow cover extent.

6. Summary and Conclusion

In this review, an overview of techniques to model and/or measure sublimation from intercepted snow, suspended snow particles, and ground snowpack was given. Previous research efforts were synthesized to further illustrate the controls on, and large variability of, sublimation in complex terrain. Challenges in estimating sublimation in complex terrain were then discussed and, despite an extensive body of literature on the topic, difficulties in measuring and modeling snowpack sublimation limit our understanding of this important water budget component, particularly on larger scales in complex terrain. Without techniques to measure sublimation at the macroscale, model estimates of sublimation cannot be validated directly. The final section of the review highlights datasets, techniques, and previous investigations that may be useful for future studies involving macroscale estimates of sublimation.

If it can be illustrated that the simulated snowpack agrees well with the seasonal evolution of SWE and simulated vapor flux agrees well with numerous previous investigations that report site specific measurements of sublimation, then model experiments have the potential to advance our understanding of sublimation at macroscales. Different model representations of canopy interception, sublimation, unloading, and sheltering may produce substantially different estimates of total macroscale sublimation in forested terrain yet sensitivity tests and ensemble simulations can increase confidence in the insight gained from model experiments that help answer important questions. How do vegetation-sublimation relationships in mountainous terrain vary by climate type, forest class (e.g., deciduous, coniferous, or mixed), healthy versus unhealthy forest, forest clearing size, forest density, intrabasin, seasonally, and interannually?

Snowpack sublimation from mountainous terrain is complex and it has been illustrated in numerous investigations that it is not a negligible component of the snow water balance. With the expectation of a continued global temperature increase, water resource issues in many mountainous regions with seasonal snow cover may develop or intensify. Therefore, advancing knowledge of all water budget components in complex terrain at the macroscale is critically important.

Conflict of Interests

The author declares that there is no conflict of interests regarding the publication of this paper.

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