VARIABILITY OF SNOW SUBLIMATION IN THE UPPER COLORADO RIVER BASIN

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ABSTRACT

Snowpack water stored in mountain environments is the primary source of water for the population of much of the western United States, and the loss of water through direct evaporation (sublimation) is a significant factor in the amount of runoff realized from snow melt. A land surface modeling study was carried out in order to quantify the temporal and spatial variability of sublimation over the Upper Colorado River basin through the use of a spatially distributed snow-evolution model known as SnowModel. Simulations relied on forcing from high resolution atmospheric analysis data from the North American Land Data Assimilation System (NLDAS). These data were used to simulate snow sublimation for several years over a 400 by 400 km domain in the Upper Colorado River Basin at a horizontal resolution of 250 m and hourly time-steps.

Results show that total volume of sublimated water from snow varies 68% or between 0.95 x 10^7 acre feet in WY 2002 to the maximum of 1.37 x 10^7 acre feet in WY 2005 within the ten years of the study period. On daily timescales sublimation was found to be episodic in nature, with short periods of enhanced sublimation followed by several days of relatively low snowpack water loss. The greatest sublimation rates of approximately 3 mm/day were found to occur in high elevation regions generally above tree line in conjunction with frequent windblown snow, while considerable contributions from canopy sublimation occurred at mid-elevations. Additional sensitivity runs accounting for reduced canopy leaf area index as a result of western pine beetle induced tree mortality were also carried out to test the models sensitivity to land surface characteristics. Results from this comparison show a near linear decrease in domain total sublimation with reduced LAI. Model performance was somewhat satisfactory, with simulations underestimating precipitation and accumulated SWE, most likely due to biases in the precipitation forcing and errors in determining precipitation phase. (KEYWORDS: snow, sublimation, snow modeling, Upper Colorado River Basin)

INTRODUCTION

Throughout much of the western United States, water reserves stored in the form of mountain snowpack provide the primary source of water for the population, agriculture and many high and middle elevation ecosystems (Doesken et al., 1996). This is particularly true in the Upper Colorado River Basin where up to 70% of annual flow originates from snowmelt alone (Christensen et al., 2007). Located in the Southwestern US in portions of Colorado, Utah, Arizona and Wyoming, the Upper Colorado River Basin (UCRB) large mountain catchment covers an area of approximately 112,000 mi^2. Seasonal runoff from this river system is heavily regulated due to the high demand for water from downstream users in California and Nevada, and to meet water export quotas for existing compacts.

The ablation of mountain snow packs through sublimation is recognized as an important factor in the removal of water throughout the winter season in mid-latitude mountain regions (Beaty, 1975, Marks et al., 1992, Pomeroy et al., 1991, 1993, MacDonald et al., 2010,). Sublimation loss can account from anywhere from 10% to 60% of the total snowpack mass, and significantly impact the water balance of the region (Schultz et al., 2004). Extreme cases of sublimation have been shown to be very efficient at removing snowpack water, with losses of up to 90% of annual snowpack on preferred alpine crests (Strasser et al., 2008), and rates exceeding 8 mm/day (Avery et al., 1992). The magnitude of sublimation has been shown to vary widely across different land surface environments and elevation (Fassnacht, 2004; Montesi et al., 2004; Molotch et al., 2007; Strasser, 2008; Fassnacht, 2010). These changes can include variability in surface features such as vegetation (Liston et al., 1995; Hiemstra et al., 2002) and topography/slope aspect (Zhang et al., 2004), as well as different environmental variables like wind, solar insolation, temperature and precipitation regime (Hood et al., 1999). Additionally, sublimation has been

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shown to vary greatly within the seasonal and sub-seasonal timeframe, with large losses during the wintertime and the potential for small amounts of condensation onto the snow surface during spring and early summer (Martinelli, 1960; Hood, 1999).

METHODS

Study Domain

The study domain was chosen to be a square area roughly centered over the UCRB covering an area of approximately 180,000 km^2 (Figure 1) and ranges in elevation from 1115 m to 4384 m. The northern and southern boundaries of the domain are defined by the Colorado state line at approximately 41.0° and 37.0° latitude, the eastern edge by the continental divide of Colorado and the western edge by the Wasatch mountain range in Utah. This domain was chosen by striking a balance between maximum areal coverage and computational resources required to carry out simulations. It encompasses the largest possible area of the UCRB, including most of the high elevation snow accumulation zones while at the same time avoiding areas that lie outside of the UCRB watershed. It is important to note that this domain excludes the Green River portion of the greater UCRB watershed, and results should not be considered representative of the entire UCRB drainage area.

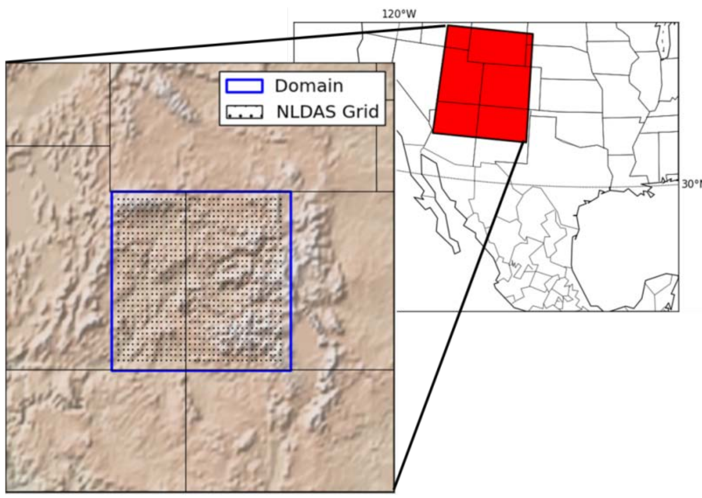

Figure 1. Location of study domain and NLDAS grid-points

Land cover, land use and vegetation vary drastically within the study domain, ranging from arid high desert environments of scrubland and short conifer forests in valley locations to dense stands of spruce and pine evergreens in the subalpine forests of the numerous mountain ranges. Precipitation during the winter is brought almost exclusively by frequent winter storms originating from the Pacific which are enhanced by orographic lifting from the high topography of the continental divide and other mountain ranges. Summertime precipitation is mostly convective in nature, but may fall as snow in the highest elevation regions well into the summer.

Model Description

SnowModel is a spatially-distributed, physically-based snow evolution model driven by input forcing fields of temperature, relative humidity, wind magnitude and direction, and precipitation (Liston et al., 2006). Snow evolution can be simulated on a range of time-steps ranging from sub-hourly to daily and on grid scales from 1 m to 1 km, and is carried out through the use of four primary sub-models. SnowModel was chosen because of its ability to simulate blowing snow sublimation, thorough documentation and computational efficiency.

Data Description

Due to the extensive area covered by the UCRB, forcing data for the snow evolution model was taken from a gridded reanalysis product rather than individual station data. Favorable validation of the NLDAS data compared to other high resolution atmospheric analysis (Cosgrove et al., 2003, Mitchell et al., 2004), combined with the ease of access granted by NLDAS, led to this data set being chosen as the primary source of surface meteorological data for the study. The North American Land Data Assimilation System (NLDAS) consists of a series of uncoupled models forced with observations and output from numerical prediction models (Cosgrove et al., 2003; Mitchell et al., 2004). Forcing data for NLDAS is generated both retrospectively and in near real-time at the National Centers for Environmental Prediction using a variety of data sources. Forcing for the non-precipitation fields are derived from the analysis fields of the North American Regional Reanalysis (NARR) that are downscaled from 32 km to the 1/8th degree (~ 14 km) NLDAS grid (Figure 1) and then temporally disaggregated to hourly time steps. The precipitation field is generated through a combination of point measurements from gauge observations and radar based precipitation estimates. that are then temporally disaggregated to an hourly time step using a combination of NWS Stage II hourly precipitation analysis (Cosgrove et al., 2003).

Elevation data were taken from the National Elevation Dataset (Gesch et al., 2009) and land cover data were taken from the 2006 National Land Cover Dataset (NLCD) (Fry et al., 2011) merged to 250 m resolution. In the case of the NLCD land cover data, a re-classification between the NLCD land cover types and the land cover types in SnowModel was required. Land cover type re-classification values are consistent with the land cover descriptions of the NLCD and SnowModel cover types including the effective Leaf Area Index (LAI) of forest land cover types found in SnowModel.

Model Configurations

While it would be possible to force SnowModel with retrospective NLDAS forcing data back to 1979, computational limitations restricted the study period to a length of 10 years. For this study the most recent 10 years of hourly forcing data from October 1, 2001 through September 30, 2011 were used. The simulation was carried out for the entire water year (WY) to avoid choosing an arbitrary end to the snow season, considering the wide range of snow free dates within the diverse environments found in the study domain. The model was then run annually for these years at resolution of 250 m and hourly time-steps. A total of 69 SNOTEL measurement locations with at least 10 years of record were chosen for validation and comparison of model output.

RESULTS

Model Results

A negligible amount of accumulated water balance error was recorded over the 10 years simulated. Domain total simulated sublimation by type over the model domain is shown in Figure 2. The annual sublimation for all types averaged 1.16 x 10^7 acre-feet of water over the ten years of simulation. The overall magnitude of total sublimation varied by 68% or between the maximum of 1.37 x 10^7 acre feet in WY 2005 and a minimum of 0.95 x 10^7 acre feet in WY 2002. The majority of the sublimation estimated by the model resulted from canopy loss, with sublimation from blowing snow only contributing a small amount to the overall amount of sublimation.

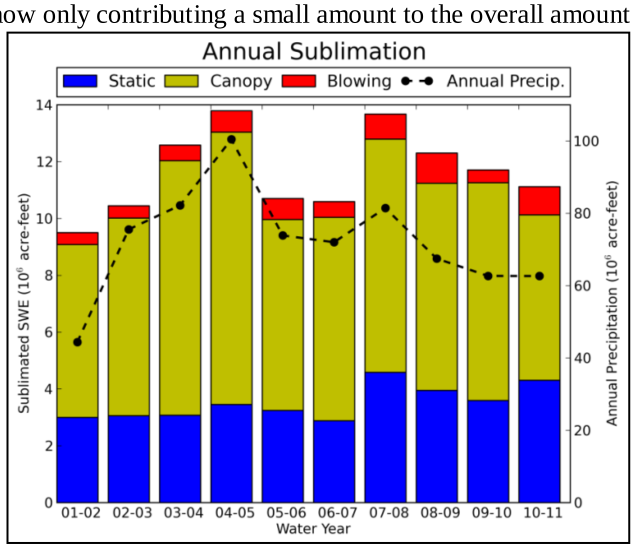

Figure 2. Annual Domain total sublimation by type from October 1, 2001 through September 30, 2011

Annual domain total sublimation for each of the components weighted by the total area over which that type of sublimation occurred during that year show that the efficiency of blowing snow sublimation rivals that of canopy sublimation, and that static surface sublimation is only about half as efficient at removing water from the snowpack as canopy or blowing snow.

Daily sublimation values from select sites show that higher rates of sublimation tend to occur during periodic episodes lasting from 2 to 5 days. Spectral analysis of daily sublimation amounts confirms this, with statistically significant peaks at the 5 and 3 day cycles. Outside of these periods of enhanced sublimation, snowpack water loss from all sublimation components generally remains less than 0.5 mm/day and lasts for several days. Average annual simulated total sublimation shows a distinct elevation-gradient, maximizing on the windward slopes of alpine regions in central and southern Colorado and minimizing in the drier valley locations. The magnitude of annual average sublimation ranges from 1-10 mm in the sheltered valleys, to isolated amounts exceeding 500 mm on preferred upwind aspects of high alpine terrain.

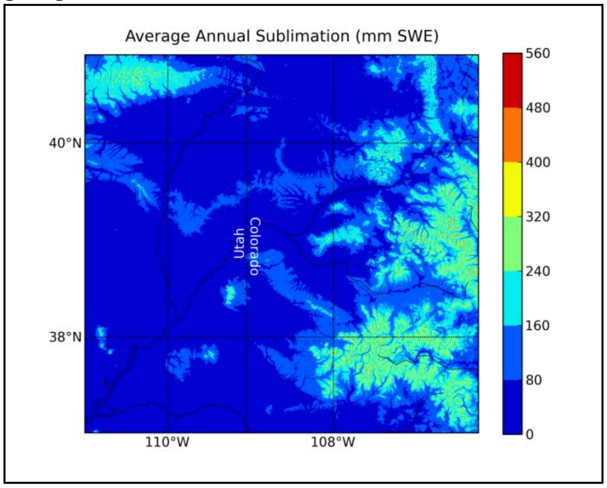

Figure 3. Average annual sublimation simulated from October 1, 2001 through September 30, 2011

Daily rates of sublimation were also computed over the entire domain using the difference in total sublimation at the end of each model day (Figure 4). Annual averages were computed by only considering days when sublimation occurred at a given grid cell. The spatial distribution of sublimation rate closely follows the distribution of total sublimation, with the highest rates located on the alpine ridgelines.

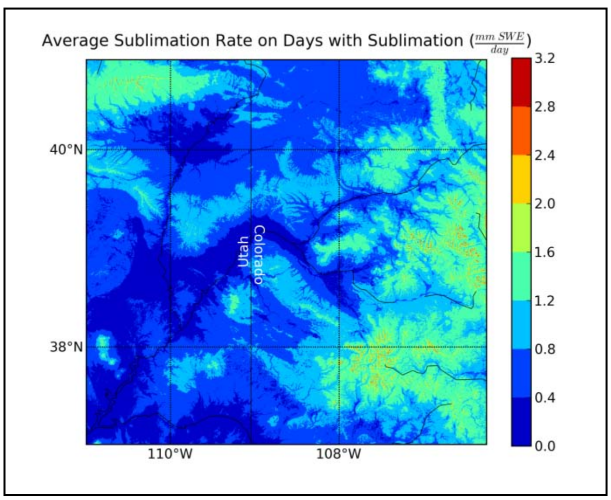

Figure 4. Average sublimation rate on days when sublimation occurred from Oct. 1, 2001 through Sept. 30, 2011

A histogram of sublimated water volume is given in Figure 5, and shows that the greatest estimates of sublimation come from lower to middle elevations in the 1300-3500 m range. Figure 5 (right) shows the same values normalized by the number of grid cells in each bin to provide sublimation per unit area. Here sublimation is seen to decrease to a minimum at the 1700 m level, then, gradually increases until the 3500 m level. Sublimation above the 3500 m level increases drastically with increasing elevation to a maximum of over 250 mm m^{-2} per year, largely due to the addition of blowing snow sublimation.

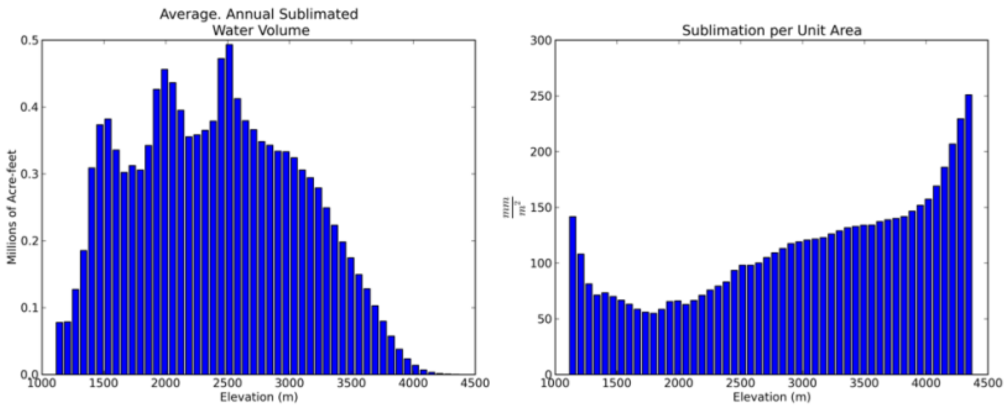

Figure 5. 10-year average of domain total sublimation (left) and domain total sublimation per unit area (right) binned by elevation

Sublimated precipitation fraction is shown in Figure 6, and ranges from 0-4% in the low valleys to 20-30% in the high mountains, with isolated areas exceeding 30% of annual precipitation. These areas of extreme sublimation loss coincide with the same areas which experience extreme daily sublimation rates.

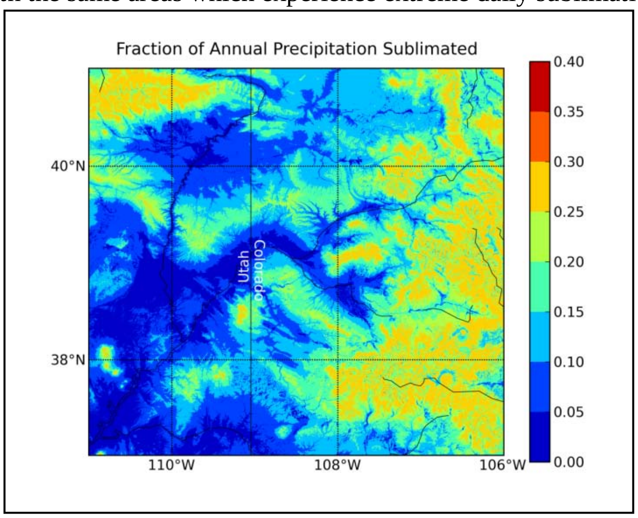

Figure 6. 10-year average of annual fraction of sublimated precipitation

Validation/Comparison with Precipitation Observations

Validation of the model results was carried out for both precipitation and accumulated SWE fields using observations collected by the Snowpack Telemetry (SNOTEL) network. It is important to note that the SNOTEL observations used in the validation were also incorporated into the precipitation forcing from the NLDAS data; however, these observations were considered the best option for comparing model output given the relative lack of consistent, long term data in the snow accumulation zones. Daily measurement values of precipitation and SWE were used, and any missing values within the observation record were discarded.

Simple least squares correlation analysis show reasonable agreement between model derived precipitation and observed precipitation, with 10 year regression coefficient of 0.65 and a correlation coefficient of 0.76 (Figure 7). Comparison between model-derived SWE values and SNOTEL observations showed a poorer relationship than the precipitation fields, with a 10 year regression coefficient of 0.38 and a correlation coefficient of 0.63 (Figure 7). Sample size for the precipitation validation was 251118, and for the SWE validation was 251100, and spanned the entire water year. Validation also appeared to be site specific, with some model grid cells consistently over or under estimating both precipitation and SWE values.

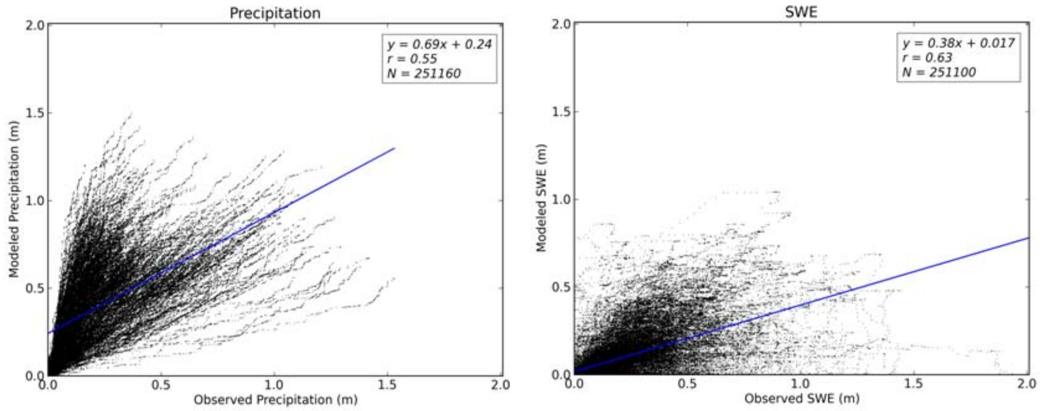

Figure 7. Comparison of observed precipitation (left) and SWE (right) from 69 SNOTEL sites to model derived values

SENSITIVITY ANALYSIS

Canopy Sensitivity

An additional simulation was carried out to test for the sensitivity of SnowModel's canopy sublimation to the LAI of the model. LAI values for the individual land types were altered following estimates made on Lodgepole Pine stands (*Pinus contorta*) impacted by the Western Mountain Pine Beetle in western North America (Pugh et al., 2012). LAI for the conifer land type was reduced by 30% and by 10% for the mixed conifer/deciduous land type. LAI was held constant for the short conifer land class because it consists of tree species that have been significantly impacted by the mountain pine beetle. A one-year simulation for WY 2004-2005 was then run using the altered values of LAI and output saved at the end of each model day.

Results from this reveal a 10% (1.01 x 10^6 acre-feet) decrease in annual canopy sublimation over the domain, with a corresponding 7% (0.26 x 10^6 acre-feet) increase in static surface sublimation compared to the control run. Changes in the amount of blowing snow sublimation were negligible (<<1%). The overall change in total sublimation for the sensitivity run is 5% (0.75 x 10^6 acre-feet) less than in the control run. Reduction in LAI also resulted in a 2% (0.10 x 10^6 acre-feet) increase in canopy unloading and a decrease of 15% in average domain canopy storage. Additional runs were made for the same water year with a 15% and 60% reduction in LAI to test the models sensitivity to various LAI values, with a near linear trend in resulting sublimation changes (Figure 8).

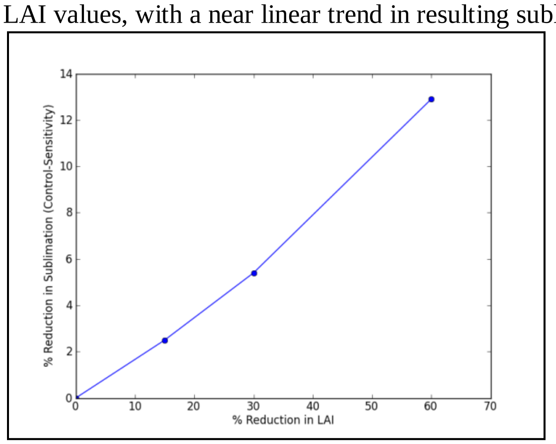

Figure 8. Percent reduction in LAI vs. percent reduction in total sublimation

DISCUSSION

Static Sublimation

The static sublimation component accounts for the smallest overall magnitude of mass flux even though it occurs over a larger land area than either blowing or canopy sublimation. Compared to blowing snow and canopy sublimation, static surface sublimation is a relatively inefficient means of sublimation due to the limited area of snow surface exposed to the atmosphere. Static sublimation is further reduced by the dense vegetation stands located over much of the lower elevation accumulation zones; however, above tree line the effects of increased ventilation are apparent, with 10 year average annual sublimation amounts exceeding 100 mm.

The effect of wind speed on static sublimation can be seen in the exchange coefficient term

$$D_e = \frac{\kappa^2 U_z}{ln\left(\frac{z}{z_0}\right)^2} \qquad [1]$$

where κ is von Karman's constant, z and z_0 are the respective observation and roughness heights and U_z is the wind speed at reference height z. Latent heat transport, and therefore sublimation, is directly proportional to wind speed at the snow surface. From this it can be seen that sublimation will occur at almost all times provided that at least some vapor pressure deficit exists and there is a non-zero wind speed.

Blowing Snow Sublimation

One of the reasons SnowModel was chosen for use in this study was its ability to explicitly simulate blowing snow processes, and the results from the SnowTran sub-model confirm the idea that blowing snow sublimation plays a significant role in the alpine snow water balance. The extreme conditions of sustained high velocity winds, intense solar radiation and large potential vapor pressure deficits found in high elevation environments leads to very efficient mass transfer from solid to vapor phase.

The influence of ventilation on sublimation can be seen in the sublimation coefficient term as (Pomeroy et al., 1991):

$$\psi(x^*, z) = \frac{1}{m}\,\frac{2\pi r\left(\frac{RH}{100} - 1\right) - \frac{S_p}{\lambda_t T_a}\left[\frac{h_s M}{R\,T_a} - 1\right]}{\frac{h_s}{\lambda_t T_a}\left[\frac{h_s M}{R\,T_a} - 1\right] + \frac{1}{D\rho_v}}\left(1.79 + 0.606\left(\frac{2rU_w}{\nu}\right)^{0.5}\right) \qquad [2]$$

This form of the sublimation coefficient shows how the rate of sublimation is proportional to the square root of the wind velocity. The influence of wind illustrates the effect that highly ventilated environments, like those found at high elevations, have on the rate of sublimation within the model where the sublimation coefficient increases rapidly with greater wind speeds.

Simulated blowing snow sublimation amounts agree with previous studies using SnowModel, where sublimation on exposed ridgelines often exceeds 500 mm annually. Because SnowTran assumes that the transport flux of blowing snow is in equilibrium with the wind field it neglects the effects of suspended snow plumes resulting from flow separation along steep ridges, a phenomenon often observed during clear, windy days on alpine peaks (Liston et al., 2007). There also remains the issue of how sublimation, especially from blowing snow, acts to modify the boundary layer through the addition of water vapor and thermodynamic feedbacks (Déry et al., 1998). For the case of SnowModel, these feedbacks have been neglected (Liston et al., 1998) and represents another source of uncertainty in estimates of sublimation.

Canopy Sublimation

The relatively high contribution of the canopy component to domain total sublimation attests to the efficiency of mass transfer of intercepted snow within the model, and is of particular interest given the widespread pine forests characteristic of the snow accumulation zones in the UCRB. Results from the canopy component of sublimation show an average loss of $5.05 \times 10^7 \frac{kg}{km^2}$, which is in-line with conservative estimates that show canopy sublimation of $4.47 \times 10^7 \frac{kg}{km^2}$ for a forested watershed in western Canada (Schmidt et al., 1992). These results show that the model simulated sublimation that were comparable to estimates made from actual observations, and reinforces the idea that sublimation returns a large portion of snowpack water to the atmosphere.

Land surface characteristics, particularly those of forests, have the ability to vary on short timescales, with extreme events such as fires resulting in changes to a large area of the surface environment in only a matter of days to weeks. In the case of the UCRB, impacts from various species of bark beetle (*Coleoptera: Scolytidae*) have led to widespread tree mortality and subsequent reduction in the canopy density which may not be represented in either the land use data or in the model parameterizations.

Changing the value of the LAI for forest land cover types will directly impact the calculated sub-canopy wind speed, $U_{subcanopy}$, which is given by

$$U_{subcanopy} = e^{\left(-(0.9\,LAI^{*})\frac{\left(1-\left(0.6\,H_{veg}\right)\right)}{H_{veg}}\right)} U_{grid} \qquad [3]$$

where LAI^{*} is the effective LAI of the forest land cover type, H_{veg} is the vegetation snow holding capacity height and U_{grid} is the interpolated wind speed. Here it can be seen that reducing the LAI will lead to an increase in the sub-canopy wind speed, and thus increase the mass transfer from solid to vapor phase. Altering the LAI will also modify the surface radiation balance by allowing more shortwave to penetrate to the surface and reduce long-wave attenuation by the canopy.

The sensitivity run of reduced LAI illustrated that changes to the forest canopy density led to corresponding changes in the amount of canopy and static surface sublimation, with a net decrease in domain total sublimation of 5%. Doubling the LAI reduction leads to an even greater reduction in sublimation, decreasing the canopy component by almost 12% from the control run (Figure 8). Even though this number is only a small fraction of the overall sublimation budget, it equates to approximately 750,000 acre-feet of water, or an equivalent 5 mm of additional SWE over the entire domain. Additional sensitivity runs show that this relationship is approximately linear with LAI reduction, and that simulated canopy sublimation is strongly dependant on the amount of snow intercepted by vegetation. In the case of the UCRB, reductions in LAI from mountain pine beetle mortality are far from homogenous in space and time, and includes tree stands in various stages of mortality and regeneration.

SnowModel uses a melt unloading scheme that assumes a constant unloading rate for above freezing. Unfortunately this parameterization does not allow for intermittent unloading events due to wind movement. The inability to explicitly simulate wind-induced unloading is desirable because unloaded snow in the low solar insolation, low wind speed and high relative humidity environment of the sub-canopy experiences far less sublimation than would snow within the canopy.

Temporal Variability

Throughout the 10 years of simulations performed, the absolute magnitude of sublimation was found to have a great deal of year to year variability, closely following the domain total precipitation. Larger sublimation amounts for years with greater precipitation is due to the larger snow covered area and longer duration of the snow-pack which allows for more mass flux, consistent with previous findings (Kattleman et al., 1991). Individual components of sublimation also show a remarkable year to year variability, particularly the blowing and canopy components of sublimation. Despite these inter-component changes, the over-all magnitude of sublimation shows no clear trend across the 10 years of simulations. The cycling between dominant sublimation components between the years is also of interest. Sublimation efficiency (e.g. the amount of water sublimated per area over which the sublimation type occurs) is dominated by the canopy during the early years, but becomes dominated by blowing snow sublimation later in the period.

The annual cycle of sublimation generally follows results from previous studies (Hood et al., 1999) that show the majority of sublimation occurs during the mid-winter snow accumulation season. It is during this time period that high wind speeds combine with low moisture content air to maximize mass flux and rapidly deplete the snowpack. The close track of daily sublimation to the snow accumulation curve illustrates the strong dependence of sublimation on available snow cover.

Sublimation peaks during the late winter and early spring period when wind speeds are greatest and average RH values begin to decline. Results show that sublimation has a tendency to occur during discrete time periods of enhanced mass flux which are then followed by corresponding periods of little or no sublimation, in line with previous work (Hood et al., 1999). Sublimation events tend to occur in cycles of about 3 to 5 days, and are followed by several days of relatively low sublimation. The magnitude of sublimation also varies greatly from event to event, with the largest events or 'sublimation storms', removing more than 10 mm of water from the snowpack over a period of a few days. This hypothesis agrees with observations by Hood et al. who note that sublimation events east of the continental divide in Colorado corresponded to down slope Chinook winds known for their dry, warm characteristics.

Spatial Variability

Results from the model simulations reveal an increase in sublimation across gradients of elevation throughout the domain. Not only do high elevation areas lose the most water from solid phase transition, but they lose it at a greater rate than low elevation areas. This characteristic is best illustrated when considering the annual sublimation bins normalized by the number of grid cells in each bin (Figure 5, right). At altitudes above 3500 m the increasing trend in sublimation becomes almost exponential, and is likely a demarcation of the typical altitude where blowing snow sublimation becomes a more dominant component of the sublimation budget by allowing for more efficient mass transfer.

Analyses of daily sublimation rate reinforces this finding, showing the highest rates of up to 3 mm/day in the high alpine regions. While such mass transfer rates appear to be quite high, they are only a third of the almost 9 mm/day reported by lysimeter measurements made in northern Arizona under clear, windy conditions (Avery, 1992). In fact, such high sublimation rates appear to be typical for mountain ranges found in desert environments, with Schultz and workers reporting rates of 3 to 5 mm/day and results from the White Mountains of California suggesting even greater rates (Beaty, 1975), indicating that the calculated rates of daily sublimation found in this study are well within the bounds of previous research. A similar pattern is found in the annual sublimated precipitation fraction (Figure 6), with the greatest loss of precipitation occurring in the highest elevations and lesser amounts in valley locations. These numbers appear reasonable compared to those found in previous studies that show between 10% and 30% of annual precipitation is returned to the atmosphere via sublimation.

This relationship of increasing sublimation with altitude has profound implications on the role that sublimation plays in the water balance of mountain environments, indicating that the greatest impact from sublimation is felt in areas with the highest concentration of snow pack water. The highly ventilated, low pressure environment of these alpine zones provides adequate driving force to efficiently transition mass from the solid to vapor phase, and also has a large reservoir of water to act upon. These efficient transfer conditions lend credence to the idea, suggested by Schmidt et al., that sublimation acts as a source of atmospheric water vapor (Schmidt et al., 1992) and significantly alter the characteristics of the atmospheric boundary layer.

MODEL PERFORMANCE

Precipitation Validation

Validation of model grid cells corresponding to the location of SNOTEL observations provided somewhat satisfactory results, with a general underestimation of precipitation by the model. Despite this shortfall on precipitation, the correlation coefficient shows reasonable agreement between precipitation in the model and in the real world with an r value of 0.76. The model appears to do better on some years than others, with a large spread in regression coefficients between individual years. Some degree of inaccuracy was anticipated in the precipitation field for a number of reasons, the most obvious being the lack of precipitation observations due to the remote and undeveloped nature inherent to the central Rocky Mountains. Radar based estimates also suffer in the rugged topography of the region, which when combined with the highly variable spatial distribution across steep elevation gradients leads to a great deal of uncertainty in the precipitation analysis; however, many of the same issues would plague station observations without the added benefit of a high temporal resolution.

SWE Validation

Validation of model derived SWE values was less than for the precipitation validation, with the model drastically under-estimating SWE accumulations across the entire domain, with substantial variability in validation from year-to-year. Correlation coefficients show a moderate relationship between the simulations and observations, indicating that snow accumulates approximately at the same time in the model as it did in the real world. Another factor that likely contributes to this poor snow-pack representation is the model's inability to properly distinguish between liquid and solid precipitation types at temperatures near freezing. SnowModel defines the transition between rain and snow when the air temperature is below 2 °C; however, the near surface air temperature may not be representative of temperatures immediately above the near surface layer, and would likely result in snow falling when the analyzed 2 meter temperature was above 2 °C.

CONCLUSIONS

Sublimation and subsequent removal of water from wintertime snow cover is a major component of the water balance for any area, and results from this study demonstrate that the magnitude and character of sublimation vary considerably across a large mountain catchment. The 10 years of snowpack simulation was carried out using the best available forcing data to quantify the change in annual sublimation magnitude. In addition, the model was run at a fine grid resolution of 250 m in order to determine the spatial characteristics of sublimation. Results from this effort indicate that the amount of sublimation varies greatly from year to year depending on precipitation amount, land cover characteristics and meteorological conditions.

Results also show significant variability in sublimation rates across gradients of elevation, with high altitude areas experiencing larger rates of sublimation due to increased wind ventilation, intense solar radiation and large vapor pressure deficits. These high sublimation rates combined with the long duration of snow cover at high elevations leads to these areas having the largest total sublimation of any location within the domain.

Based on the results of this study, the author concludes that

1. Sublimation is a major component of the water balance within the UCRB, and results in a significant loss of snowpack water
2. Sublimation generally increases at higher elevations, with a sharp increase in sublimation above 3500 m MSL
3. Model derived sublimation is most efficient when snow is blowing or saltating
4. The magnitude of sublimation varies greatly on inter-annual timescales
5. On daily time scales, sublimation appears periodic in nature, with 'events' of enhanced sublimation resulting substantial loss of water from the snowpack

Furthermore, these sublimation events are driven by periods of extremely dry, and most importantly windy, conditions that are sustained for several hours or a few days.

Future Considerations

Of particular interest is the response of sublimation to changes to the forest canopy in conjunction with the ongoing bark beetle infestation. The resulting net decrease in over-all sublimation found in the sensitivity run illustrates that even subtle differences in the land surface can have profound implications on the water balance. This investigation only considered short term effects of tree mortality, namely the reduction of LAI due to needle loss; however, the future forests of the UCRB will likely see even more drastic changes as dead trees begin to fall allowing for a much different make-up of stem heights, tree species and ground cover.

Finally, while the NLDAS data used to force the simulations is believed to be the best for use over such a large domain, the relatively poor performance of the model in accurately simulating both precipitation amount and especially SWE amount shows that precipitation fields could be improved. In addition to improving precipitation estimation, more work needs to be done on how precipitation phase is determined. If this study were to be carried out again, it should be done in a manner that puts less emphasis on spatial extent in order to focus more on small scale processes, that hold the most influence over sublimation. These considerations should include:

1. Most accurate representation of land cover type possible, including explicitly simulating vegetation processes such as wind unloading
2. Increased resolution to capture fine scale blowing snow processes
3. Improved representation of snow cover, particularly focused on using better precipitation forcing

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