THE EFFECT OF VARIABLE PATTERNS OF SNOW DEPOSITION AND DRIFTING ON SNOWMELT, RUNOFF, AND STREAM DISCHARGE IN A SEMI-ARID MOUNTAIN BASIN

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ABSTRACT

In semi-arid mountainous regions, local topography and canopy cover strongly affect wind patterns during storms which alters snow distribution and causes the development of hydrologically significant drifts. While large drifts may cover only 5–15% of the basin area, they can hold 50% or more of the basin SWE at peak accumulation and 75–100% of the basin SWE in late spring. Snowmelt from the drifts provides essentially the only source of water to the basin in late spring and early summer, and therefore directly affects basin ecology, runoff response, and the basin hydrograph. To understand how snow distribution and drifting affect the timing and magnitude of snowmelt and the delivery of meltwater to the stream, we simulate both the patterns of snow deposition and melt over a small headwater basin in the Owyhee Mts. (the Reynolds Mountain East basin (0.36 km^2)) using the energy balance snowmelt model ISNOBAL. Simulations were run for several snow seasons in the 1980's for which time-series aerial photos monitoring of the location and depletion of snow drifts during melt-out were available. Precipitation input was modeled as a function of topographic exposure relative to storm-event winds and the difference in storm-event catch between a sheltered site and an exposed site. Snowcover in the modeled drift zones lasted well into the spring and showed good agreement with drift assessments made from time-series late-season aerial photography. This experiment shows that disparate patterns of snow deposition and melt, including the effect of drifting, which is typical of semi-arid mountain basins, can be modeled as a function of terrain and vegetation.

INTRODUCTION

In mountainous basins, local topography and canopy cover strongly affect snow distribution, snowpack energy fluxes, and resultant melt rates. In regions with little topographic variability it has been demonstrated that canopy cover and its effect upon snow surface energy exchanges is the dominant control on snow distribution [e.g. Link and Marks, 1999]. However, it has often been cited that in mountainous regions wind is the dominant control on snow distribution [e.g. Elder et al, 1991; Luce et al., 1998]. Though much has been accomplished to effectively model the effect of canopy cover on radiation fluxes and snow interception [Marks et al., 1998, 1999; Link and Marks, 1999a,b; Marks and Winstral, 2001], an efficient means of accounting for wind effects on snow distribution in alpine terrain is only beginning to be tested [Marks et al., 2001]. Within the energy balance snowmelt model, ISNOBAL [Marks et al., 1999], we spatially varied precipitation inputs and wind speeds using terrain-based parameters representative of wind exposure [Winstral, 1999], and accounted for the effects of canopy cover upon radiation fluxes using methods established by Link and Marks [1999] to model snow accumulation and melt for the 1986, 1987, and 1989 snow seasons.

Reynolds Creek Experimental Watershed

The Reynolds Creek Experimental Watershed (RCEW) is a 233 km^2 basin, located in the Owyhee Mountains in Southwestern Idaho. Annual precipitation is concentrated in wintertime snows over the high elevation southern extent of the basin. At the watershed scale, downstream irrigation is heavily reliant upon melt of the winter snowpack, while at the sub-basin scale, alpine ecosystems are linked to intra-basin snow accumulation differences, particularly snowdrifts.

Study Site

The Reynolds Mountain East sub-basin (RME) is a 0.36 km^2 basin with an elevation range of 2027 - 2137 m. Hydrology of the sub-basin is snowmelt driven producing an average 583 mm of runoff per year, compared to 75 mm of runoff for the RCEW as a whole. Two climate stations representing the extremes of exposure in RME were used to develop the distributed climate fields for this study. Site 176, the "ridge site", has an unobstructed fetch for several km to the south (the prevailing wind direction). Site RMSP, the "grove site", is a typical SNOTEL (snow telemetry) site located just in the lee of site 176. The 40-year average annual precipitation at site 176 is 553 mm, while at site RMSP, it is 848 mm.

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OBJECTIVE

Account for the effects of wind and canopy cover in a distributed model of snow accumulation and melt in a sub-alpine first-order watershed.

PROCEDURE

Snow accumulation and melt was modeled from October 1 through June 30 for water years 1986, 1987, and 1989 using the distributed energy balance snowmelt model, ISNOBAL. Climate forcing data for ISNOBAL consisted of meteorological data from two sites within the study area that were distributed based on elevation, topographic shelter, and canopy cover at an hourly time-step.

Spatial validation data consisted of a time series of snow extent maps classified from aerial photographs taken during the meltout phase. Additionally, daily modeled surface water inputs were compared to the basin hydrograph.

Climate Summary

The 1986 and 1989 winter seasons were both wetter than average with basin runoff 152% and 128% of average respectively. The 1987 winter season (basin runoff 40% of average) was the fifth driest in the 33-year record.

Energy Balance Snowmelt Model

ISNOBAL is a two layer coupled energy and mass-balance model, designed to be applied over a DEM grid where climate, precipitation, and snowcover data are available. The model assumes the snowcover is composed of two layers that are homogeneous in temperature, density, and liquid water content. The surface layer is a fixed thickness, and the lower layer represents the rest of the snowcover. The model is designed to be driven by standard meteorological data such as that available from typical monitoring sites.

Forcing Variables

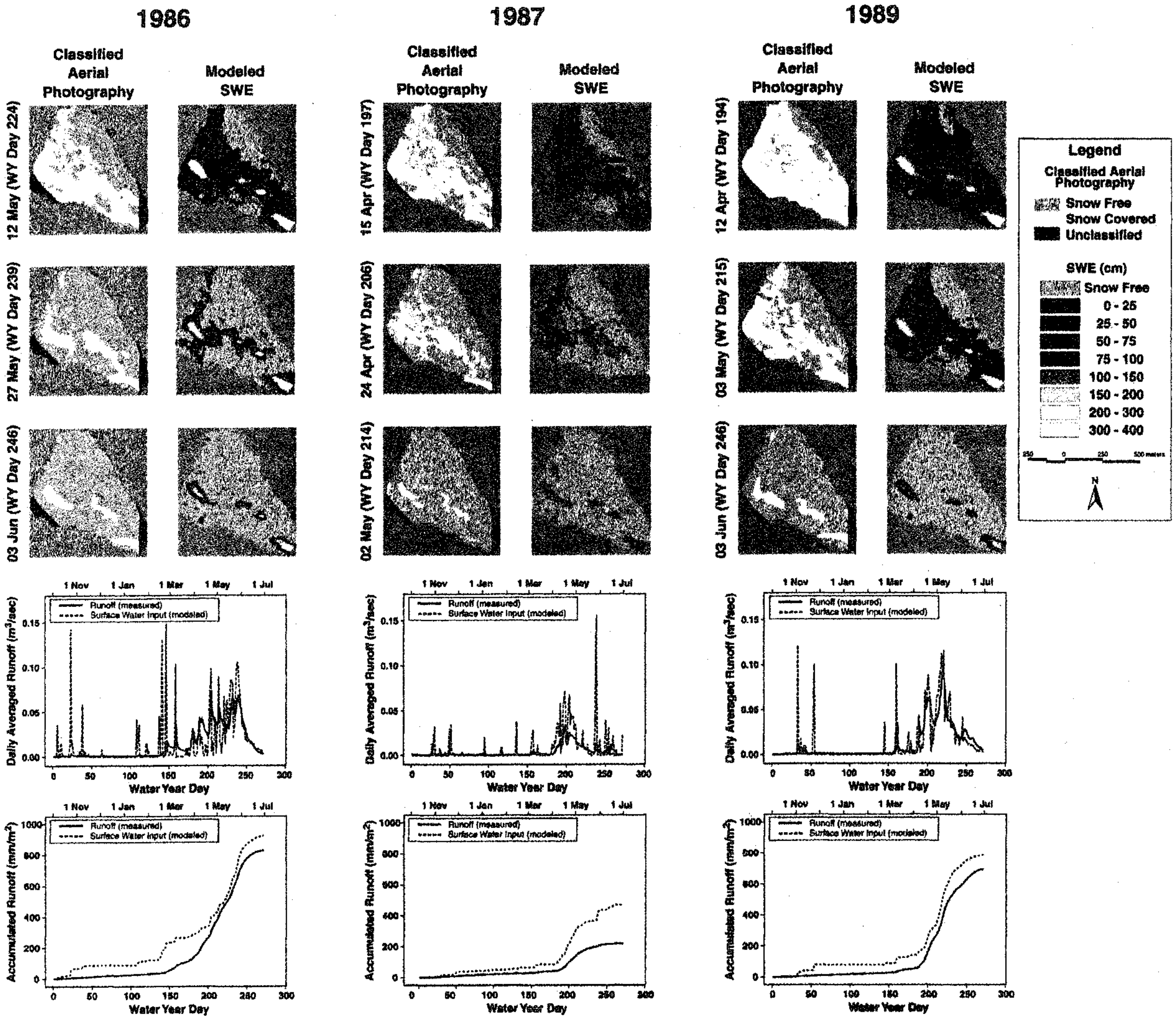
All variables were modeled over the 10 m DEM grid. The grid containing the RME sub-drainage was 88 x 80 cells with the sub-drainage itself consisting of 3762 grid cells. Air temperature, vapor pressure, and soil temperature were distributed over the DEM using data from climate stations and a simple linear distribution with elevation. Wind speed and precipitation were corrected for topographic structure (i.e. shelter/exposure). Drift zones relative to specific wind directions were determined using the maximum upwind slope (MAXUS) and upwind slope break (TOPOBREAK) parameters, respectively [Winstral, 1999]. MAXUS values were subsequently modified for the sheltering effects of canopy cover. Libraries of exposure and drift images were created for all possible wind directions in 5° increments. The library images were recalculated whenever an additional 1 m of snow had accumulated in a drift zone with modeled snow depth added to the DEM elevation. Minimum and maximum MAXUS values were set equal to the respective values at the ridge and grove sites. Hourly wind data from the two climate stations were interpolated over the appropriate wind-factor field to derive hourly wind speed images. Hourly precipitation data from the two stations were similarly draped over the wind-factor field to derive hourly images of precipitation distribution. For each storm, precipitation in the drift zones was determined as a function of the ratio of total storm precipitation catch at the grove site to that at the ridge site and hourly accumulation at the grove site.

Net solar radiation and thermal radiation were calculated using topographically corrected clear-sky values (Marks and Dozier, 1979) corrected for cloud cover and canopy effects (Link and Marks, 1999). Cloud cover reduction factor was computed from measured solar radiation. Solar radiation transmission and absorption were computed by splitting cloud-corrected radiation into visible and infrared, and beam and diffuse streams. Diffuse radiation transfer was computed by applying a hemispherical view factor for each canopy class. Beam radiation transfer was computed using Beer's Law and applying a canopy height and extinction coefficient for each class (Price, 1997). Thermal radiation computed by applying a hemispherical view factor, and assuming canopy temperature equal to air temperature.

RESULTS

Simulations were done over the RME sub-basin for the 1986, 1987, and 1989 water years. The model was initiated on October 1 in each of the three water years, simulating both the development and ablation of the seasonal snowcover. Simulated SWE closely matches measured SWE at the snow pillow site (the grove site). As shown in Fig. 1, aerial photographs from each of the three years were used to verify the simulated patterns of snow cover, showing good correspondence between simulated and measured snow covered area (SCA) throughout the melt season. Simulated runoff (snowmelt plus rain) is also compared to measured runoff from the drainage during melt. With the exception of a few spring rain events, simulated runoff closely matches measured during all three years. Cumulative simulated and measured runoff show good agreement during 1986 and 1989, but diverge during 1987.

Figure 1:



CONCLUSIONS

Using topographic structure as a surrogate for wind effects, and established canopy/radiation relationships we were able to efficiently and accurately model snow accumulation and melt over the study basin. The modeled snow distribution corresponded well with the pattern observed in the time-series of aerial photographs taken during the meltout phase in all three years. The modeled surface water inputs were consistent with the hydrologic response measured at the basin outlet. Hydrograph peaks directly matched the timing and relative magnitudes of modeled surface water inputs during meltout. During the snow accumulation phase, when most meltwater contributes to soil moisture recharge, very little response was observed at the basin outlet. In 1986 and 1989, once snowmelt intensified and the soil became saturated, modeled water inputs closely matched the hydrograph. In 1987, a well below-average snow year, hydrologic response appears dampened relative to the modeled inputs. We believe this to be the product of soil water contents being below field capacity.

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