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The influence of the spatial distribution of snow on basin-averaged snowmelt

Charles H. Luce,[1]* David G. Tarboton[2] and Keith R. Cooley[3]

[1]*USDA Forest Service, Rocky Mountain Research Station, 316 E. Myrtle, Boise, ID 83702, USA,*
[2]*Department of Civil and Environmental Engineering, Utah State University, Logan, UT 84322-4110, USA,*
[3]*USDA Agricultural Research Service, Northwest Watershed Research Center, 800 Park Blvd., Plaza 4, Suite 105, Boise, ID 83712-7716, USA*

Abstract:

Spatial variability in snow accumulation and melt owing to topographic effects on solar radiation, snow drifting, air temperature and precipitation is important in determining the timing of snowmelt releases. Precipitation and temperature effects related to topography affect snowpack variability at large scales and are generally included in models of hydrology in mountainous terrain. The effects of spatial variability in drifting and solar input are generally included only in distributed models at small scales. Previous research has demonstrated that snowpack patterns are not well reproduced when topography and drifting are ignored, implying that larger scale representations that ignore drifting could be in error. Detailed measurements of the spatial distribution of snow water equivalence within a small, intensively studied, 26-ha watershed were used to validate a spatially distributed snowmelt model. These observations and model output were then compared to basin-averaged snowmelt rates from a single-point representation of the basin, a two-region representation that captures some of the variability in drifting and aspect and a model with distributed terrain but uniform drift. The model comparisons demonstrate that the lumped, single-point representation and distributed terrain with uniform drift both yielded poor simulations of the basin-averaged surface water input rate. The two-point representation was a slight improvement, but the late season melt required for the observed stream-flow was not simulated because the deepest drifts were not represented. These results imply that representing the effects of subgrid variability of snow drifting is equally or more important than representing subgrid variability in solar radiation.

KEY WORDS spatial variability; distributed catchment modelling; snow hydrology

INTRODUCTION

The spatial variability of snowmelt processes has received increasing attention in recent years (Blöschl *et al.*, 1991; Kirnbauer *et al.*, 1994). Varying precipitation input, drifting and solar radiation intensity on sloping surfaces all relate to topography and contribute to the heterogeneity of surface water input from snowmelt (Seyfried and Wilcox, 1995; Tarboton *et al.*, 1995). One of the more marked effects of spatially variable accumulation and melt is the effect on the timing of snowpack releases.

While the importance of topography in determining snow accumulation and melt has been well established, methods to represent the effects of topography on drifting have not been well explored. Several

* Correspondence to: Charles H. Luce, USDA Forest Service, Rocky Mountain Research Station, 316 E. Myrtle, Boise, ID 83702, USA. e-mail: cluce@rmci.net

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researchers have examined the detailed physics of snow transport under known wind fields (Tabler, 1975; Tabler and Schmidt, 1986; Pomeroy and Gray, 1995). Others have approached the problem through empirical means (Elder *et al.*, 1989, 1991; Blöschl and Kirnbauer, 1992). Jackson (1994) and Tarboton *et al.* (1995) estimated drifting for a small watershed by calibrating a drifting parameter in an energy and mass balance snowmelt model at each grid cell. While this calibration appears to be stable for the years at the site for which it was done, the relationships between topography and drifting are not easily generalized. If the distributed drifting cannot be calculated based on readily obtained spatial data, it becomes another of the unknown or unknowable parameters in distributed models discussed by Beven (1996).

Because precise mapping of a drifting parameter may be difficult, a general characterization of the effect through a subgrid parameterization for a larger scale model may be more manageable. At 30 m grid resolution, drifting can be explicitly represented; for larger model elements, only the net effect of drifting needs to be described. This study addresses the question of what level of detail is necessary in representing topography and spatial variability of snow drifting in distributed snowmelt modelling.

METHODS

Study area

The study was carried out using data from the Upper Sheep Creek subbasin of the Reynolds Creek Experimental Watershed in south-western Idaho, which enjoys a long and rich history of hydrological research (Stephenson and Freeze, 1974; Cooley, 1988; Duffy *et al.*, 1991; Flerchinger *et al.*, 1992, 1994; Jackson, 1994; Seyfried and Wilcox, 1995; Tarboton *et al.*, 1995, amongst others). Much of the work has focused on runoff generation mechanisms in the basin, concluding that groundwater flow through layered basalts is the primary source of stream flow. All of the above studies have noted the importance of the snowdrift that forms in the south-west portion of the basin in contributing water during the period of greatest runoff. This background of previous work measuring snow drifting (Cooley, 1988), a previously developed and calibrated distributed hydrological model (Jackson, 1994; Tarboton *et al.*, 1995) and an understanding of the basin's hydrology provide a good foundation from which to explore the effects of the spatial distribution of snow on basin-averaged snowmelt.

The Upper Sheep Creek watershed has an area of 26 ha, with elevations between 1840 and 2040 m (Figure 1). Low sagebrush communities cover the north-east portion of the basin, and big sagebrush communities cover most of the south-western half of the basin. Aspen grow in a narrow strip along the north-east-facing slope where the drifts form (Figure 2). Severe winter weather and winds keep the aspen dwarfed to heights between 4 and 7 m. Average annual precipitation is 508 mm, and the first-order stream exiting the basin is ephemeral.

Study outline

We used distributed and lumped snowpack models to examine the ability of simplified representations of spatial variability in topography and drifting to estimate surface water input. Each of four simulations was considered as a hypothesis and compared with distributed snow water equivalent measurements, such as those described by Cooley (1988), and the timing of basin outflow through a weir. First, the fully distributed snowmelt model was run with the distributed topography and distributed drift factors reported in Jackson (1994) and Tarboton *et al.* (1995). This simulation was used to check the validity of the snowmelt model and to calculate the basin-averaged snowmelt flux for conditions approximating the actual conditions. The next two simulations were simplifications of that representation. From aerial photography and field observations, it is clear that drifting occurs primarily on shadowed, north-east-facing slopes, while sunnier, south-west-facing slopes are scoured by prevailing winds. This strong covariance in processes yields a shallow snowpack over time on south-west-facing slopes versus a deep snowpack over time on north-east-facing slopes and suggests that a division of the basin into a north basin and a south basin may yield some of the observed

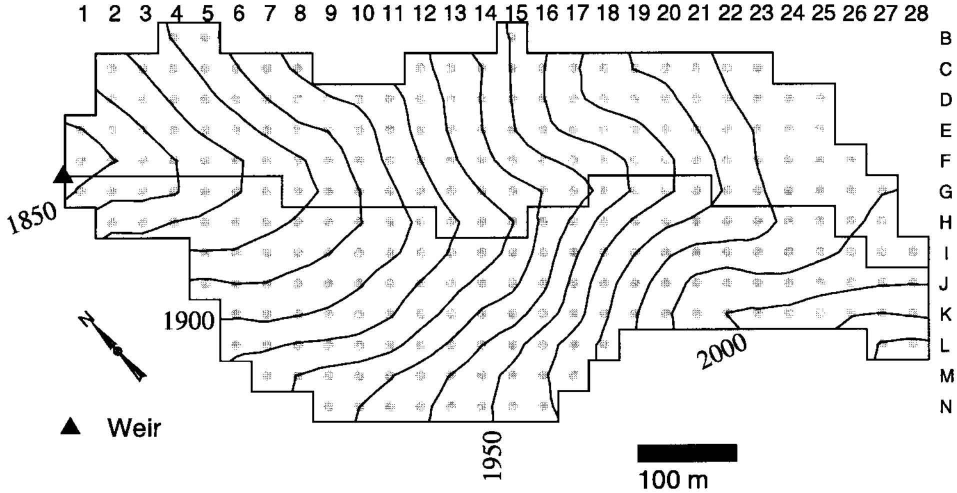

Figure 1. Map of Upper Sheep Creek with snow survey grid. Contour interval is 10 m. The area above the line separating the two halves of the watershed will be referred to as the north-east side later in the paper. The area below the line will be referred to as the south-west side

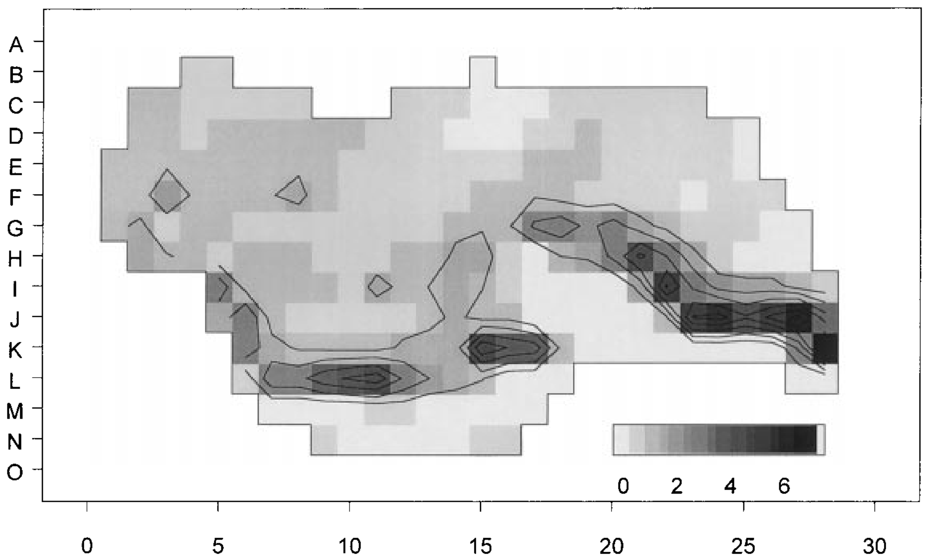

Figure 2. Map of drift multipliers used at Upper Sheep Creek (after Jackson, 1994)

basin-wide behaviour in snowpack distribution. This single division into two regions is a substantial simplification compared with the 255 cells used in the fully distributed model.

The second simplification treated the basin as a single unit with a single aspect, slope and drift factor. This simulation was run to confirm that it gave a poor approximation to the data and to see where the two-region simplification fitted between the fully lumped and fully distributed representations.

A final simulation examined the importance of the drift factor. In this simulation, the spatial variation in topography was preserved, but the first factor was set to unity everywhere, removing the spatial variability resulting from drifting, but modelling the control that topography has over incident radiation.

Data collection

Measurements of snow water equivalent were taken on nine dates in 1993 with a snow tube and scale. A grid guided distributed sampling over the watershed (Figure 1). The spacing on the grid is 30·48 m (100 ft), and the long axis is oriented 48 degrees west of north. Precipitation, temperature, relative humidity and incoming solar radiation were measured for water year 1993 at a weather station near location J10. Wind speed was measured at D3. Flow is measured at a weir at location F0.

Model description

The snowmelt model is an energy and mass balance model with a vertically lumped representation of the snowpack. It is more completely described in Tarboton and Luce (1997). Two primary state variables are maintained in the model: snow water equivalent, W(m), and internal energy of the snowmelt and top 40 cm of soil, U(kJ m^{-2}). U is zero when the snowmelt is at 0 °C and contains no liquid water. These two state variables are updated according to

$$\mathrm{d}U/\mathrm{d}t = Q_{\mathrm{sn}} + Q_{\mathrm{li}} - Q_{\mathrm{le}} + Q_{\mathrm{p}} + Q_{\mathrm{g}} + Q_{\mathrm{h}} + Q_{\mathrm{e}} - Q_{\mathrm{m}}$$

$$\mathrm{dW}/\mathrm{d}t = P_{\mathrm{r}} + P_{\mathrm{s}} - M_{\mathrm{r}} - E$$

where Q_{sn} is net solar radiation, Q_{li} is incoming long-wave radiation, Q_{le} is outgoing long-wave radiation, Q_{p} is advected heat from precipitation, Q_{g} is ground heat flux, Q_{h} is the sensible heat flux, Q_{e} is the latent heat flux, Q_{m} is heat advected with melt water, P_{r} is the rate of precipitation as rain, P_{s} is the rate of precipitation as now, M_{r} is the melt rate and E is the sublimation rate. The model is driven by inputs of precipitation, air temperature, humidity, wind speed and incoming solar radiation. Snow surface temperature, a key variable in calculating latent and sensible heat fluxes and outgoing long-wave radiation, is calculated from the energy balance at the surface of the snowpack, where incoming and outgoing fluxes must match. These simulations were run on a 6-hour time-step.

The effect of plant canopy on snowmelt is parameterized by decreasing the albedo of the snow surface as the snow depth decreases below the canopy height. This parameterization is most appropriate for short vegetation, such as sagebrush. Because the aspens are free of leaves until the soil warms slightly, errors introduced by not considering the taller canopy are minimal.

The distributed model runs the point model (described in the preceding two paragraphs) at each cell in the grid (Figure 1). The model uses a drift multiplier to estimate enhancement of local incoming snow through wind transport. The fraction of precipitation falling as rain or snow is a function of temperature. The fraction falling as snow is multiplied by the drift multiplier to estimate grid cell precipitation. The drift multiplier was calibrated from 1986 snow survey data from Upper Sheep Creek. Drift multipliers were adjusted at each grid cell to match the snow water equivalent on 25 February and 26 March 1986 (Jackson, 1994; Tarboton *et al.*, 1995). Values of the multiplier over the basin are shown in Figure 2 (Jackson, 1994) and ranged from 0·2 to 6·8, with an average of 0·975. A value less than one indicates that the basin loses more snow to neighbouring basins than it gains.

Distributed solar radiation was estimated based on pyranometer data at the weather station, which was used to calculate an effective atmospheric transmission factor. Local horizons, slope and azimuth were used to find local sunrise and sunset times and to integrate solar radiation received on the slope during each time-step. The calculated atmospheric transmission factor characterized cloudiness for incoming long-wave radiation calculations.

Site characteristics used for the single-point and two-point representations of the basin are summarized in Table I. For the single-point model, the average basin elevation and drift factor were used. Slope and aspect were calculated along the long axis of the basin to estimate the lumped basin behaviour. For the two-point model, representative cells were picked for the north-east and south-west sides of the basin to set slope, aspect and elevation. Each point was assigned an average drift factor for the region it represented.

RESULTS

Maps of observed snow water equivalent over Upper Sheep Creek watershed are shown in Figure 3a. The effect of drifting in concentrating snow, and, consequently, late season snow water equivalent along the south-west side of the basin is evident. Maps of modelled snow water equivalent with the fully distributed snowmelt model (Figure 3b) show a generally similar pattern. Table II lists the basin-averaged snow water equivalent from the observations and the model, showing that the fully distributed model tends to overestimate snow water equivalent in the late melt. Plotting observed against modelled data for each date (Figure 4) shows that the fully distributed model overestimates snow water equivalent for locations with moderate to high snow water equivalents, but underestimates snow cover where there is little snow, with systematic overestimation most apparent in the early melt season. The correlation coefficient (Pearson's r) and a measure of fit to the 1 : 1 line (Wilmott, 1981; Wilmott *et al*., 1985) are given in Table III. It should be

Table I. Effective site characteristics for single-point and two-point representations of the basin

	Single-point representation	North-east side	South-west side
Slope	0·159	0·286	0·345
Aspect (°)	312	299	357
Drift factor	0·975	0·62	1·29
Elevation (m)	1925	1912	1939
Relative area (%)	100	47	53

Table II. Basin-averaged snow water equivalent (m) from observations and models

Date	Observed	Model with drift	Model no drift
10 February	0·22	0·28	0·28
3 March	0·28	0·38	0·39
23 March	0·23	0·23	0·10
8 April	0·18	0·16	0·00
15 April	0·17	0·16	0·00
29 April	0·13	0·13	0·00
12 May	0·09	0·07	0·00
19 May	0·04	0·03	0·00
25 May	0·02	0·01	0·00

Table III. Agreement between modelled and measured images

Date	Pearson's r	Willmott's d
10 February	0·83	0·90
3 March	0·84	0·90
23 March	0·90	0·94
8 April	0·88	0·93
15 April	0·89	0·94
29 April	0·89	0·94
12 May	0·90	0·94
19 May	0·87	0·92
25 May	0·65	0·76

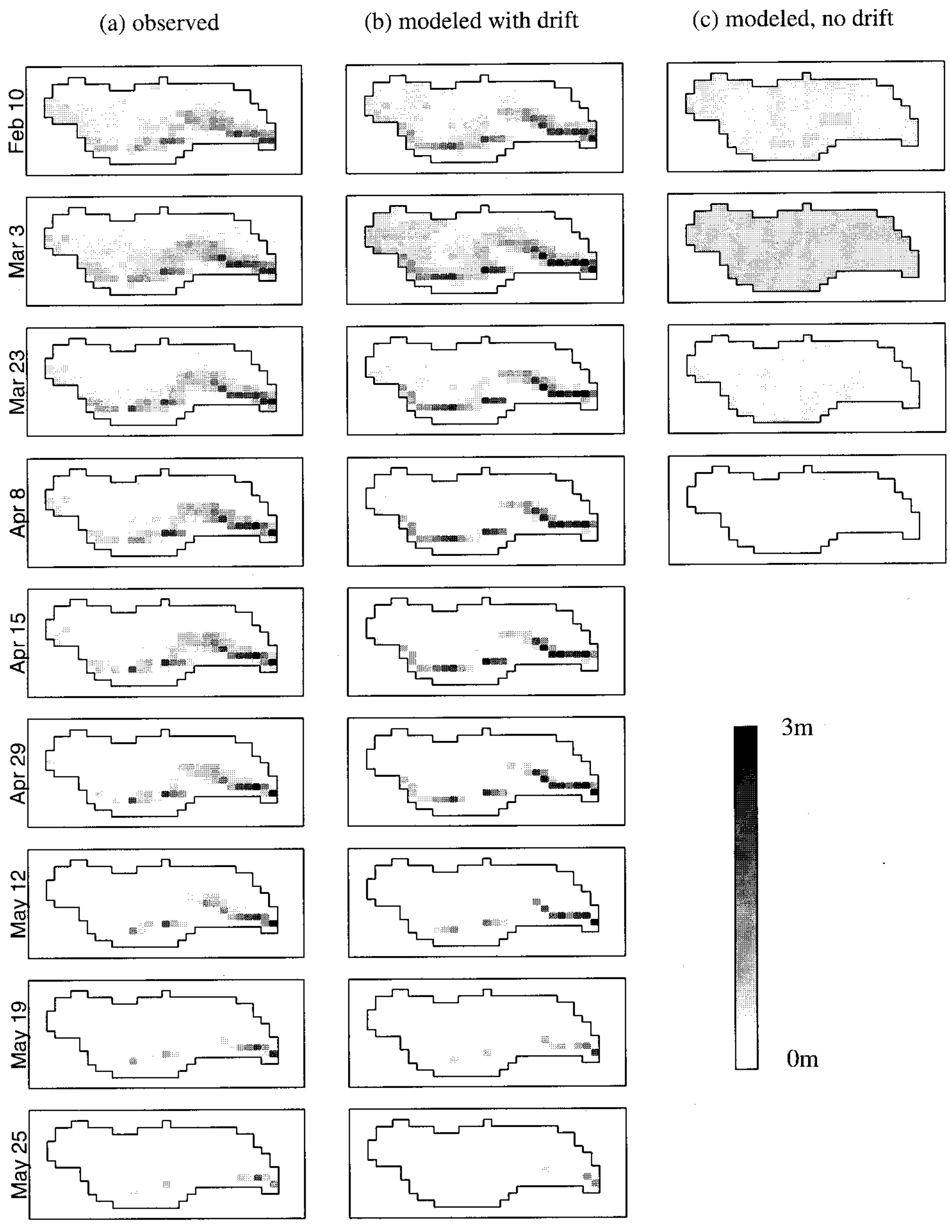

Figure 3. Snow water equivalent mapped over the basin on nine dates of snow survey in 1993 for: (a) observed, (b) modelled with spatially varying drift factor and (c) modelled with uniform drift factor. No snow modelled after 8 April with uniform drift factor

noted that there is a degree of spatial autocorrelation, the structure of which is not known exactly. The goodness-of-fit implied by the r values may therefore be somewhat overstated. These results, obtained with multipliers calibrated using 1986 measurements (Jackson, 1994), show drifting patterns that compare favourably with 1993 observations, suggesting consistency in drifting from year to year.

A comparison of the predictions of the distributed model using a uniform drift factor over the basin (Figure 3c) with the observed data (Figure 3a), shows that drifting is an important process in creating variability in snow water equivalence across the basin and in determining the timing of melt outflows. Differences

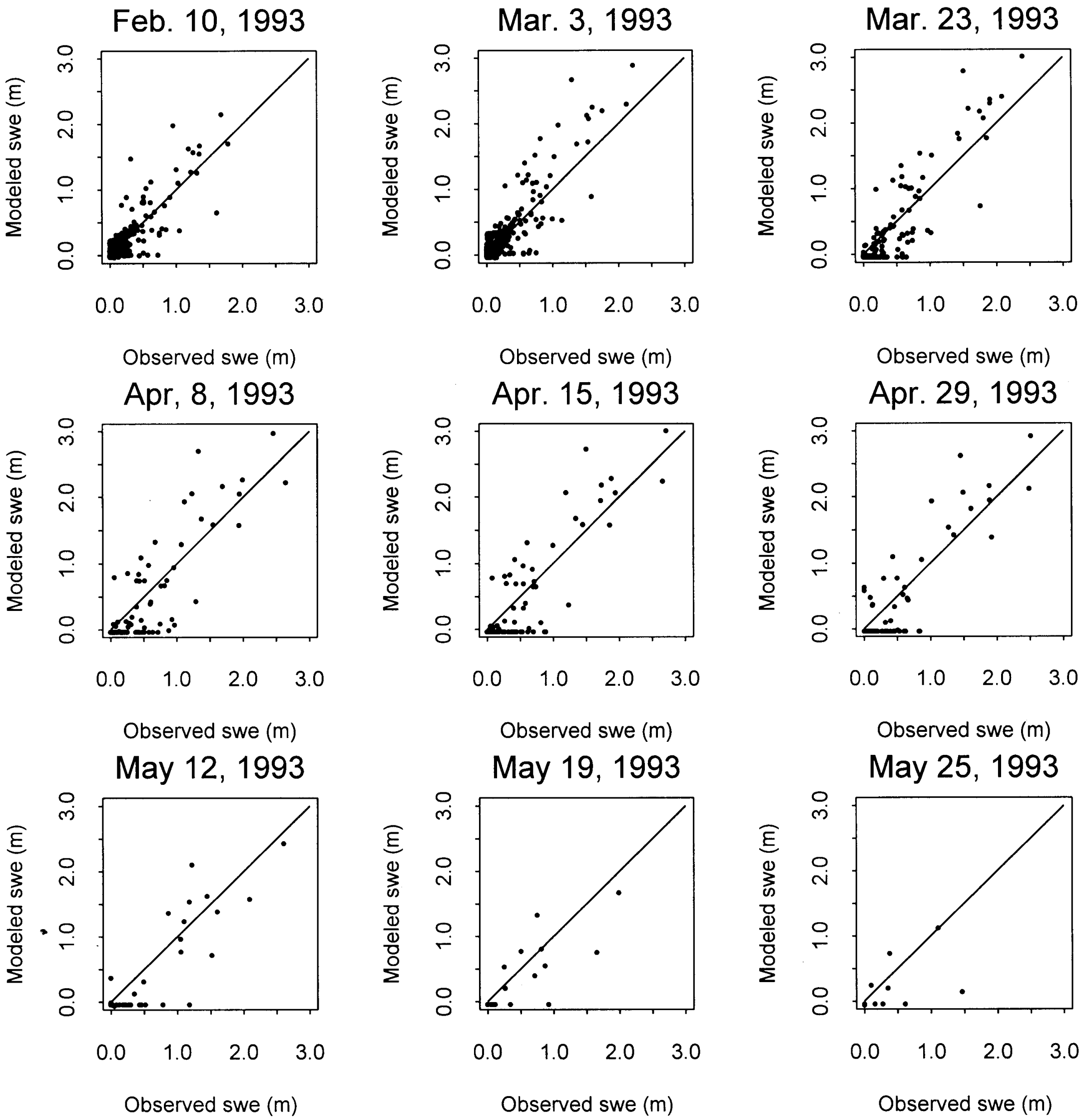

Figure 4. Comparison of observed and modelled snow water equivalent for each snow survey date. The line through each plot is the 1 : 1 line

in melt caused by differences in solar radiation and temperature across the basin are not great enough to explain the spatial patterns of snow water equivalent values over the basin. The snow water equivalent modelled in this manner shows considerably less variability than that measured or modelled with a spatially varying drift factor. Consequently, all cells in the basin become snow-free almost simultaneously, and the persistence of the snowpack in the basin is dramatically reduced relative to observations. This result implies that spatial variability in snow drifting has a greater effect on the behaviour of Upper Sheep Creek than spatial variability in solar radiation and temperature.

As an additional check on the behaviour of the fully distributed model, we also compared modelled and measured surface water inputs (snowmelt plus rain) averaged over the period between snow water equivalent measurements. Cumulative surface water input and sublimation from the snowpack (loss) can be calculated

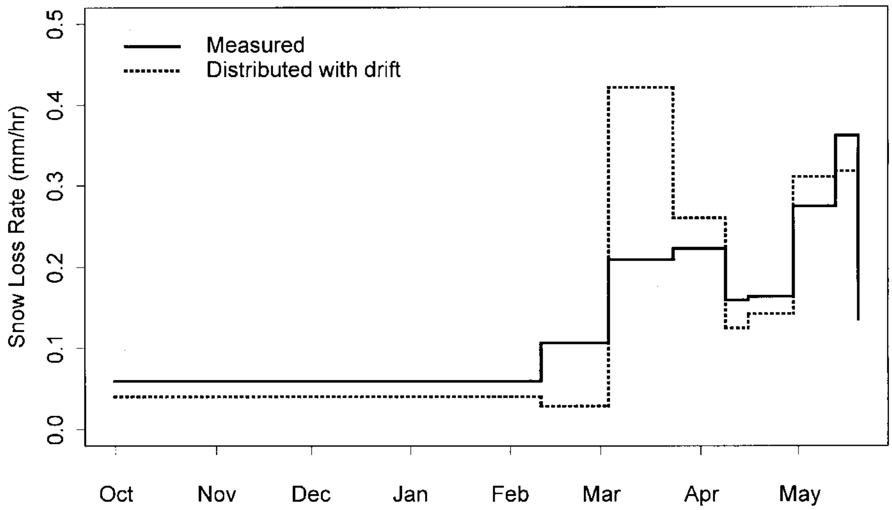

Figure 5. Measured average loss rate and modelled loss rate over time. Losses are the sum of melt and sublimation

as the measured cumulative precipitation less the measured snow water equivalent on a particular date. The average loss rate for the periods between measurements was then calculated from the cumulative values. Figure 5 shows the average snow loss rate based on the measurements and on the fully distributed model plotted over time. Before the melt season begins, the measured rates are slightly greater. During the second measurement interval (10 February to 3 March), the fully distributed model lost less snow, which increased the error in snow water equivalent seen on 3 March in Table II. During the next measurement interval, the model overpredicted losses, mostly as melt. From Figure 3, it appears that much of the difference is in the south-facing part of the basin, which has low snow water equivalents. The model shows build-up and loss of snow in this area, while the measurements indicate that perhaps no accumulation occurred.

Figure 6 shows the calculated basin-averaged surface water input rate versus modelled basin-averaged surface water input rate for the four models (fully distributed, single-point, two-regions, distributed without drifting). To prepare Figure 6, we subtracted the modelled sublimation from the measured loss rate to estimate the 'measured' surface water input rate. Because sublimation is small relative to melt during the melt season, this is a very small correction. A striking feature of Figure 6 is how well the distributed model with drifting performs, except for one measurement period (3–23 March), where the surface water input is substantially overpredicted. The other models show poor comparisons between measured and modelled surface water input rates. These results suggest that the basin-averaged surface water input rates from the fully distributed model, which includes snow drifting, are reasonably representative of the actual surface water input rates experienced by the basin during the late melt season. They also show that the alternative models considered in this study give poor predictions of melt outflow rates.

Stream flow is a second source of evidence that can be used qualitatively to come to the same conclusion that, of the four models examined, only the fully distributed model gives reasonable estimates of snowpack outflow. Flerchinger *et al.* (1992) provide a conceptual model of runoff generation in the basin, suggesting that early melt primarily serves to recharge the groundwater, while later melt generates stream flow through a groundwater response. During average snow years (such as 1993), they found response times through a confined aquifer of the order of 3–5 days. From the cumulative modelled surface water inputs over water year 1993 and the cumulative stream flow (Figure 7), it can be seen that the timing of basin-averaged surface water input rates for the three simplified models differs from that of the fully distributed model. Very little surface water input (flat line) is predicted by the simplified models during the period of greatest stream flow (steep line), while the fully distributed model is still predicting substantial outflow during that time period

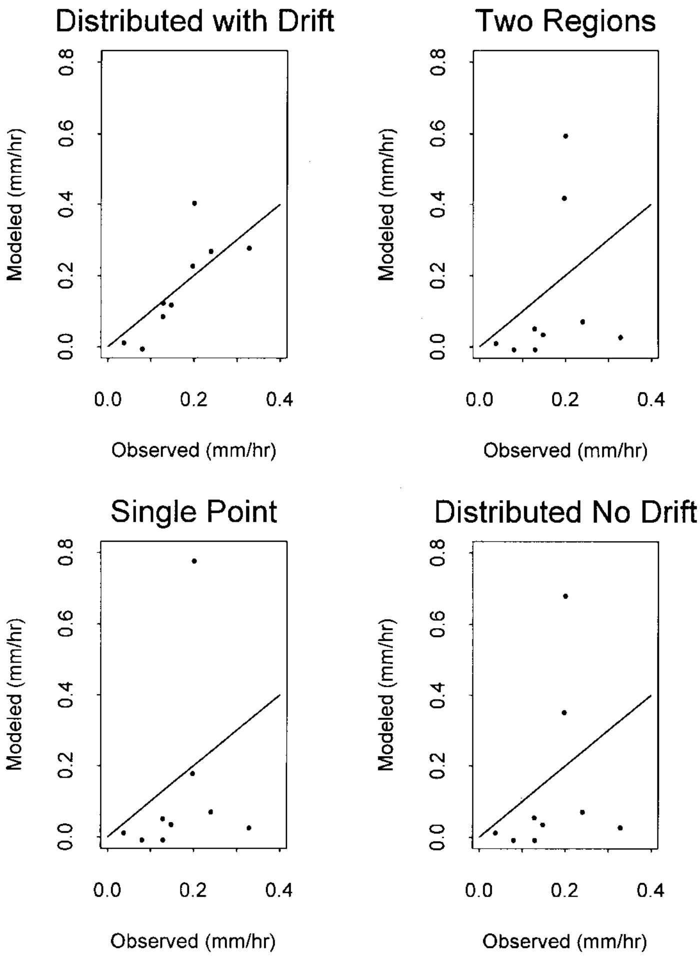

Figure 6. Observed and modelled average surface water input rate for the nine periods defined by the nine snow water equivalent measurements and the beginning of the water year (no snow)

(steep line). Timing is a little easier to compare precisely in Figure 8, from which the same conclusion may be drawn. Comparison with Figure 7 shows that the brief spikes in surface water input predicted by the three simplified models after mid-April (rainfall) represent little water. The fully distributed model is the only model that predicts significant melt late in the season, coinciding in timing with the observed rise of the stream flow hydrograph. The other models show surface water inputs concentrated almost entirely in the month of March, which is an unlikely source of water for peak stream flow in May.

DISCUSSION AND CONCLUSIONS

In semi-arid mountainous watersheds, such as Upper Sheep Creek, wind plays a large role in redistributing snow, and the spatial variability and pattern of snow water equivalent is highly dependent on wind-induced drifting. Deep snowdrifts provide melt water into late spring. Using detailed snow water equivalent measurements and distributed snowpack modelling, we examined the effects of spatial variability of snow accumulations on snowmelt processes at the scale of a small watershed (~400 m across). We found that representing basin snowmelt as a single point yields inaccurate results. Using two regions with contrasting

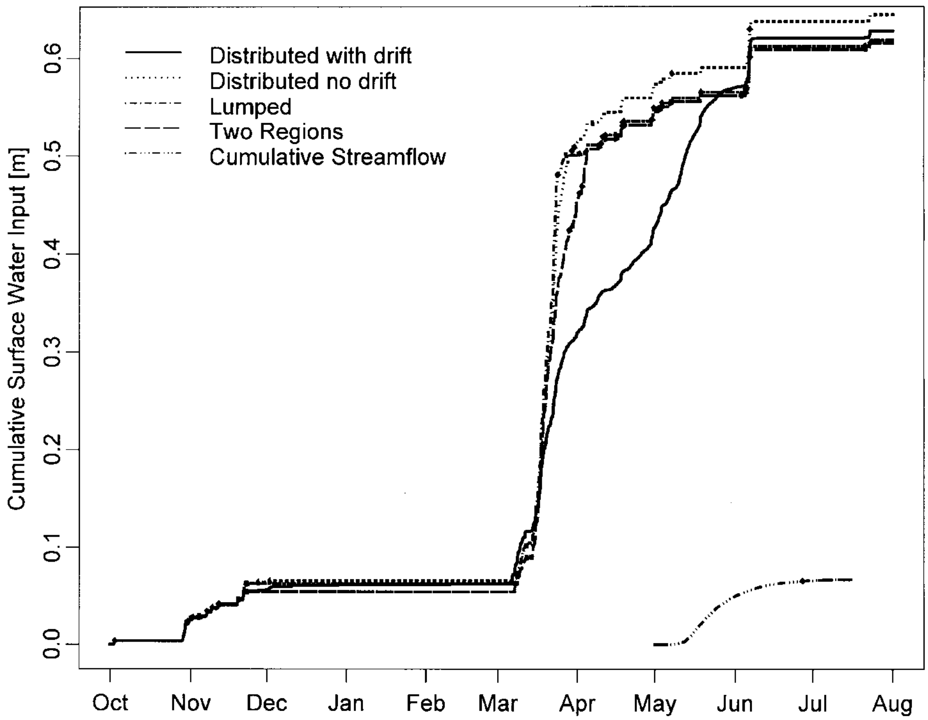

Figure 7. Cumulative surface water input for each of the four models and cumulative stream flow for the period October 1992 to July 1993

drifting and solar input to represent the basin improves the simulations little. We also examined the relative contribution of solar input and drifting to the observed spatial patterns of snow water equivalent and the temporal patterns of surface water input. Our results show that detailed snow drifting information, which may be difficult to obtain, is equally or perhaps more important than modelling the effects of local topography on radiation.

This examination relied heavily on a distributed snowpack model for reference, and some effort has been made to test how appropriate the model is for this basin. Comparisons of measured and modelled patterns of snow water equivalent on each measurement date showed reasonable agreement. The model showed some bias towards overestimation of snow water equivalent in the early melt season, with better agreement in the middle and late melt season. Surface water input was slightly underestimated throughout the accumulation season, and overestimated in the early melt season. By mid to late melt season, there is generally better agreement in surface water input rates. From the maps in Figure 3, it appears that much of the discrepancy centres on the south-west-facing slope. Because of the generally low snow water equivalents on these slopes, it is likely that the calibration using 1986 data resulted in poor estimates of the drift factor. Alternatively, the drifting here may have been inconsistent between the 1986 and 1993 snow seasons. This source of error demonstrates how sensitive the timing of basin snowmelt is to estimates of distributed drifting and how difficult those estimates are to obtain. Between the time the snow on the south-west-facing slope melted and the end of May, the distributed model snow water equivalent and melt rates compared favourably with measurements. During this time period, melt from the much deeper drifts on the north-east-facing slopes contributed most of the surface water inputs. These deeper drifts on the north-east-facing slopes contributed most of the surface water inputs. These deeper drifts are caused by prevailing winds and are probably much more consistent from year to year, as indicated by the noted correlation between vegetation patterns and drift patterns in Upper Sheep Creek (Flerchinger *et al.*, 1994).

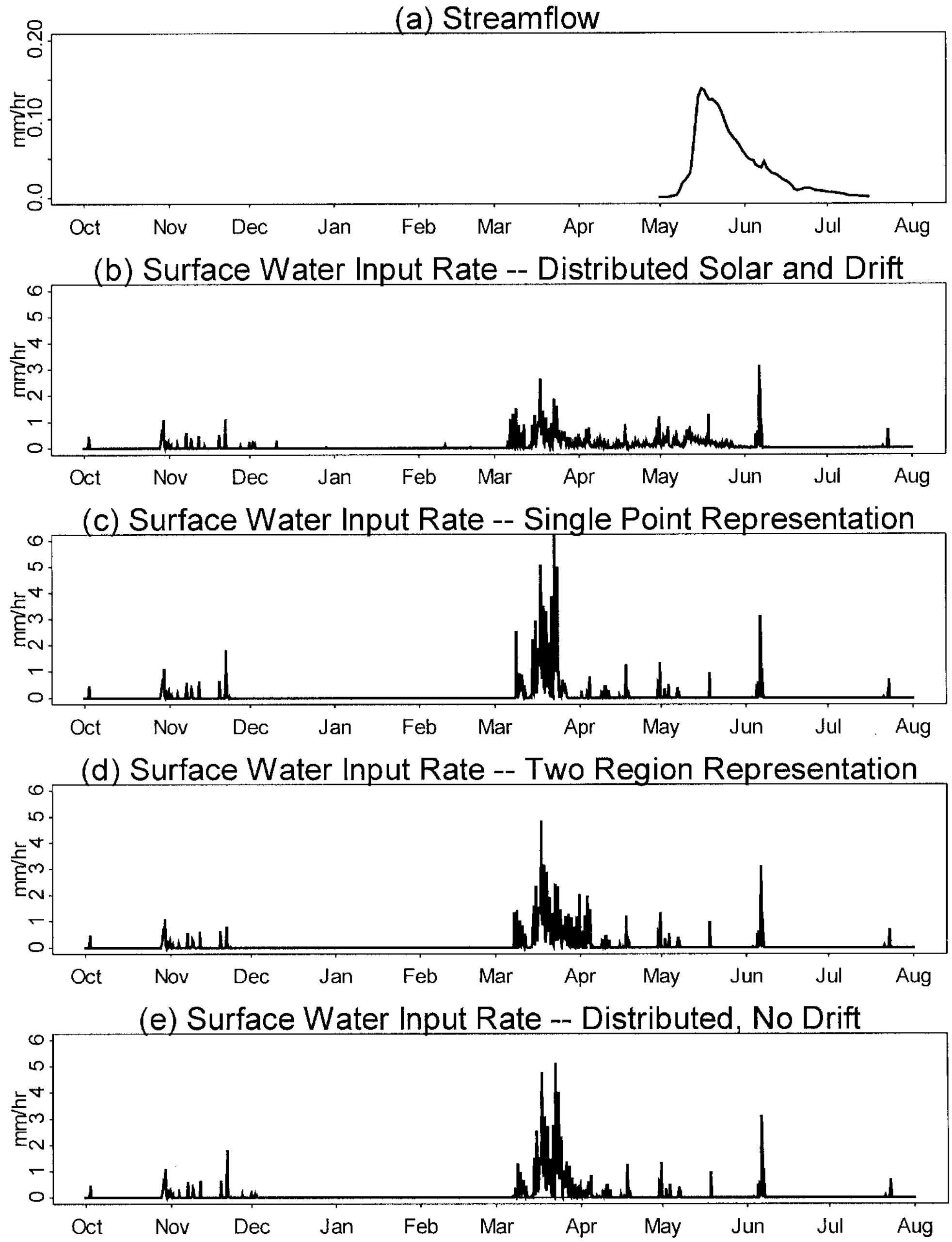

Figure 8. Surface water inputs from snowmelt and basin outflow for the period October 1992 to July 1993. (a) Observed stream flow, (b) distributed model with drift multipliers, (c) lumped model, (d) two-region model and (e) distributed model with no drift multipliers

Snowmelt modelling at the catchment scale is generally done as part of water balance modelling. There is some question as to whether potential errors in drifting, such as those found in the distributed snowmelt model, would propagate through to runoff generation estimates. In Figure 8, the fully distributed model shows a peak basin-wide surface water input rate during March. From Figure 5, we know that the surface water input rates predicted by the fully distributed model in Figure 8 are about twice what they should be during March. Because runoff in this basin occurs from sustained input to the small portion of the basin under the largest drifts (Stephenson and Freeze, 1974), it is unlikely that the relatively small depth of melt modelled on the south-west-facing slope, where the errors appeared to be the greatest, would appear as runoff. Errors in this area of the basin would most likely be manifested as errors in evapotranspiration. The concentrated surface input under the drifts, up to 3 m over the melt season, yields most of the runoff through subsurface flow (Flerchinger *et al*., 1992) and saturation overland flow, suggesting that errors in the amount of snow drifting over the area of the deep drifts could be translated directly into errors in runoff.

Accurate estimates of snow drifting in a basin are difficult to obtain, but they are important to the prediction of basin snowmelt. Errors in estimates of both basin-wide evapotranspiration and basin runoff may occur from errors in estimates of snow drifting when using a distributed hydrology model. An integrated or 'lumped' representation of the basin snowpack is one way to avoid this problem when only basin-averaged information is desired. None of the lumped representations used in this study included a subgrid parameterization that adequately represented the effects of drifting, and consequently none provided a reasonable simulation of meltwater inputs to the basin. An important challenge lies in finding a subgrid parameterization that addresses this important source of spatial variability in snow water equivalent. A bigger challenge may lie in relating the basin-wide surface water input rate to runoff generation processes that are spatially dependent on drifting.

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