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Distributed Snowmelt Simulations in an Alpine Catchment
1. Model Evaluation on the Basis of Snow Cover Patterns

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This paper presents an attempt at deterministically modeling spatially distributed snowmelt in an alpine catchment. The basin is 9.4 km^2 in area and elevations range from 1900 to 3050 m above sea level. The model makes use of digital terrain data with 25 m grid spacing. Energy balance components are calculated for each grid element taking topographic variations of solar radiation into account. For each grid element albedo and snow surface temperatures are simulated. Model performance is evaluated on the basis of snow cover depletion patterns as derived from weekly air photographs. The use of spatially distributed data allows for addressing individual model components. Results indicate that the basic model assumptions are realistic. Model inadequacies are shown to arise from processes not included in the model such as avalanching and long wave emission from surrounding terrain as well as inaccurate model parameters.

1. INTRODUCTION

Snowmelt runoff forecasts are needed for many purposes including flood warning, reservoir management and hydrochemical problem identification. Traditionally, snowmelt modeling has been governed by the operational need for forecasts. Most operational models are spatially lumped. Although these models, generally, give good results in terms of prediction capabilities [*World Meteorological Organization (WMO)*, 1986] there are indications that they do not adequately represent the underlying physical processes [*Golding*, 1974; *WMO*, 1986; *Braun*, 1988]. Typically, errors of the snowmelt routine are transferred to the runoff model or vice versa when the calibration is based on runoff. For some purposes the use of unrealistic models can be misleading. Such applications include ungauged catchments and land use and climate change. They also include the simulation of extreme situations. These are regarded as hydrologically changed conditions as compared to the average situation where data are available. *Dozier* [1987] pointed out that the increased understanding of snow science has not yet been translated into more realistic runoff models and emphasized the need for distributed snowmelt models.

Much work has been done on snowmelt models at the site or hillslope scale [e.g., *Dunne et al.*, 1976; *Price and Dunne*, 1976; *Obled and Rosse*, 1977; *Male and Granger*, 1981; *Jordan*, 1983*a*, *b*; *Morris*, 1983; *Akan*, 1984; *Morris*, 1989]. Numerous papers have been published on distributed model components such as radiation [e.g., *Dozier*, 1980; *Olyphant*, 1986] and some papers on the distribution of water equivalent [*Woo et al.*, 1983*a*; *Elder et al.*, 1989]. However, no more than a few studies deal with spatially distributed snowmelt models. *Charbonneau et al.* [1981] presented a model which accounted for variations in shortwave radiation and snow surface temperature at slopes of different aspect. However, the benefits of using the complex model for runoff simulations could not be shown. Therefore, *Obled and Harder* [1979] dealt with analyzing the processes to be included in a high-relief snowmelt model, rather than with modeling exercises. A fundamentally different approach of model evaluation was taken by *Blöschl et al.* [1989] and *Leavesley and Stannard* [1989, 1990] by verifying a distributed model with spatially distributed snow cover data. This paper follows that line.

This paper and a companion paper present an attempt at deterministically modeling spatially distributed snowmelt in an alpine catchment. This paper deals with model evaluation on the basis of snow cover patterns. Specifically, the objective was to address individual processes using different evaluation schemes. The companion paper is concerned with model predictions and is oriented toward operational applications.

It is recognized that for most engineering purposes runoff is needed whereas this paper is restricted to the simulation of melt rates only. Therefore, this paper is viewed primarily as a contribution to expand our understanding of snowmelt processes in alpine terrain. It is hoped to provide also some impetus for the development of more realistic snowmelt runoff models.

2. STUDY AREA AND MEASUREMENTS

The study was conducted in the Längental catchment, Tirol, at 47°12′N, 11°E, in the Austrian Alps (Figure 1). The basin is 9.4 km^2 in area and elevations range from 1900 to 3050 m above sea level. Topographically, the basin consists of two major units. The lower part comprises east and west-facing slopes including talus fans with typical slopes of 35° to 40°. The upper part is open to the east. The southeast corner of the basin is formed by three prominent cirques. Most of the catchment lies above the timber line. The annual precipitation averages about 1200 mm, 50% of which falls as snow. In the lower parts of the catchment the snow cover period typically lasts from November to May, whereas the upper parts become bare in July.

This analysis focuses on the 1989 snowmelt season. Due to the low winter precipitation the basin water equivalent was lower than normal. However, during the ablation period frequent precipitation occurred; on average rain or snowfall was recorded on two out of three days. The ground was frequently covered by a shallow snowpack which quickly disappeared.

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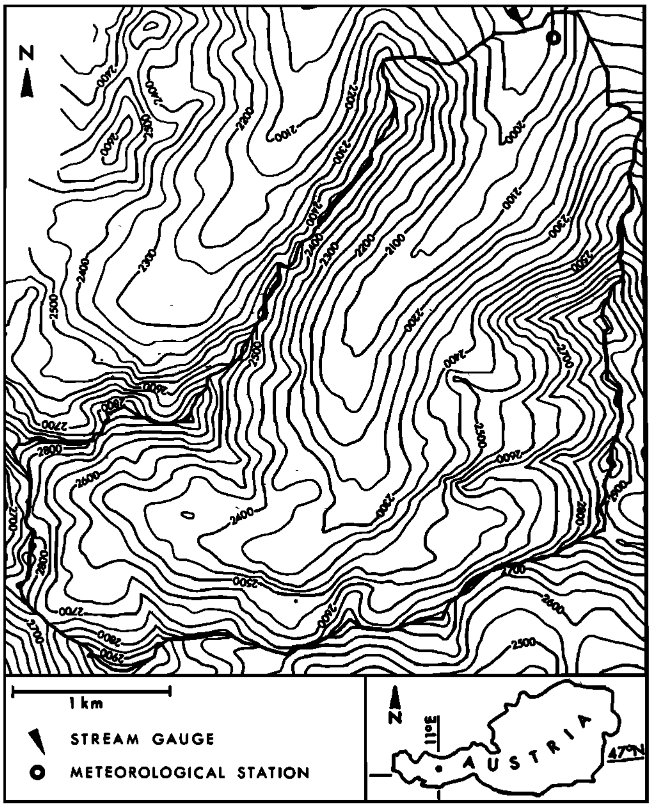

Fig. 1. Map view of the Längental basin, Tirol; contour interval 50 m.

Atmospheric data used for the analysis include global radiation, air temperature, humidity, wind speed and precipitation on an hourly basis. These variables were observed at the Kühtai station (1930 m above sea level) near the basin outlet [see *Kirnbauer and Blöschl*, 1990]. Cloudiness was determined from frequent visual observations and plots of radiation components were used in interpolating between observations, particularly during the night. Stream discharge at the outlet of the watershed was measured with a stage recorder and a twin overflow weir [*Moschen*, 1990]. Additionally, air temperatures at Finstertal (2330 m above sea level, 700 m east of the catchment boundary) were used (also see *Blöschl* [1991]).

A field program was undertaken in late April to assess the distribution of water equivalent in the basin. The selection of the sites was based on typical terrain types as outlined by *Woo et al.* [1983*a*, *b*]. These included different elevations, slopes and aspects. Measurements were designed to be representative of an area of roughly 50 × 50 m each. This was accomplished by numerous snow depth measurements over that area and a few density profiles.

Snow cover patterns were mapped on the basis of oblique aerial photos nine times during the 1989 ablation period. Snow cover boundary lines were manually identified in the printed photos and transferred to the map scale by digital monoplotting [*Radwan and Makarovic*, 1980]. Subsequently, vector data were rasterized to yield 25 × 25 m square ground elements. A single element was allowed to be either snow covered or snow free. *Hochstöger* [1989] and *Blöschl and Kirnbauer* [1991*b*] provide more explicit information on the methods used for snow cover mapping.

3. THE MODEL

The snowmelt model presented here is conceptual and distributed with space. Parameters were externally derived from the literature or from field observations.

The model is based on digital terrain data with 25 m grid spacing. Accumulation and melt are simulated for each grid element and snow cover conditions are assumed to be homogeneous over an element. The grid size is believed to be of an adequate scale to represent the rough topography of the basin [*Wood et al.*, 1988]. The variability of processes with a length scale larger than the grid size is addressed explicitly, whereas the subscale processes are assumed to be implicitly parameterized, i.e., their integral effect is examined. Therefore, the model representing processes in a single element should not be termed a "point model" since it is inherently linked to the size of that element. This becomes obvious, for example, when the snow cover in an element partly disappears.

Although there have been some attempts at deterministically modeling topoclimate [e.g., *Tesche and McNally*, 1988], here, the spatial extrapolation of meteorological variables is based on very simple assumptions. Air temperature is assumed to decrease linearly with altitude based on the readings at two stations. No positive gradients are allowed to avoid unrealistic temperatures at high elevations. *Blöschl* [1991] showed that little uncertainty is introduced by the above assumptions. Wind speed and relative humidity are taken as invariant across the basin and precipitation is assumed to increase by 30%/km [see *Kuhn and Pellet*, 1989]. In the case of snowfall, deposited snow is corrected for redistribution effects due to wind and gravity. This correction scheme is analogous to the interpolation procedure of water equivalent described below.

Energy balance components are calculated for each grid element on an hourly basis. Global radiation measured at Kühtai is divided into a diffuse and a direct component on the basis of cloudiness [*Neuwirth*, 1982]. The separation is based on a linear relation between the ratio of diffuse to global radiation and cloudiness, which is adjusted with data from overcast days and from clear sky days before sunrise

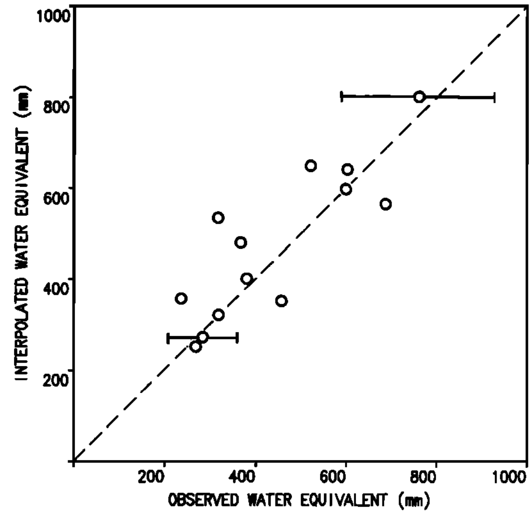

Fig. 2. Observed snow water equivalents versus predictions of interpolation scheme, late April 1989. Bars indicate standard deviations within a 50 × 50 m area.

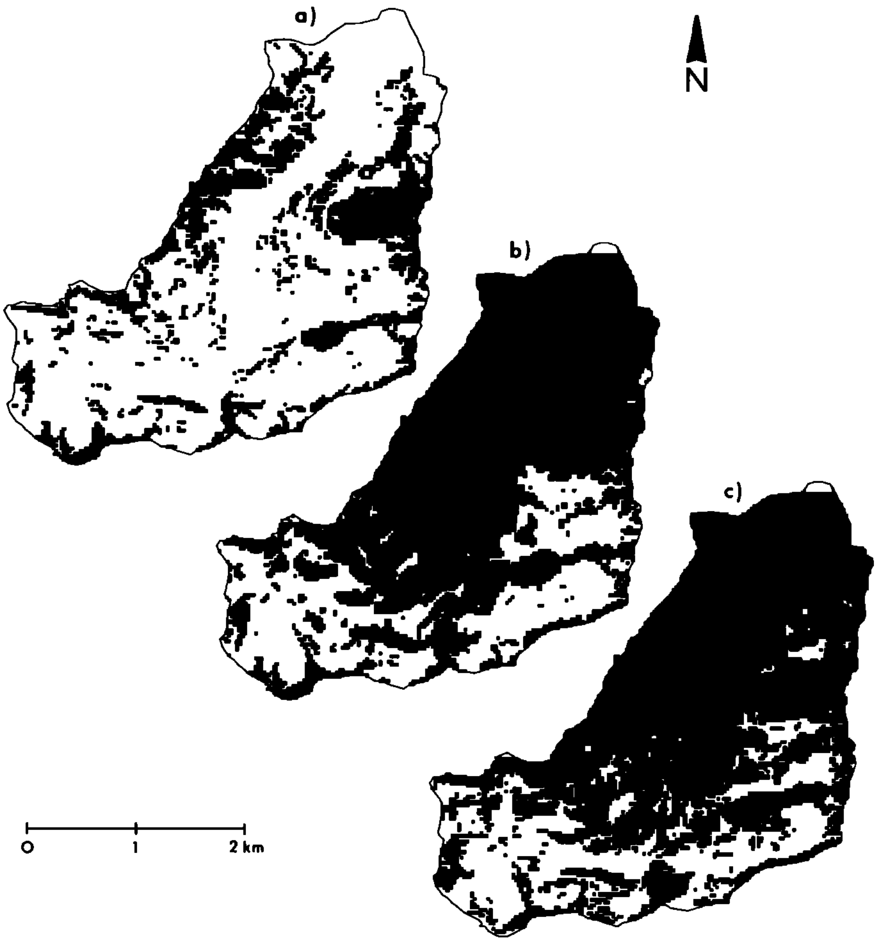

Fig. 3. (*a*) Observed snow cover April 24, 1989 (initial condition); (*b*) simulated snow cover June 26, 1989; and (*c*) observed snow cover June 26, 1989. Solid areas denote bare ground and open areas denote snow cover.

(assuming 100% diffuse radiation in both cases). The computation of direct radiation for a tilted plane including shading by surrounding terrain is based on simple geometric principles [*Obled and Harder*, 1979]. Diffuse radiation is considered to be isotropic over the sky dome disregarding shading effects. This is a good approximation for overcast skies and high albedoes when the sky and the surrounding snow cover have radiatively similar behavior. Errors induced in the clear sky case are discussed by *Obled and Harder* [1979]. Shortwave reflection from adjacent terrain is approximated by the assumption of a tilted plane surrounded by flat, reflecting terrain [*Obled and Harder*, 1979]. Shortwave albedo is assumed to vary solely with the age of the snow surface [*U.S. Army Corps of Engineers*, 1956; *Blöschl*, 1991]. It is recognized that this is a very simple approach (e.g., disregarding variations with changes in solar zenith angle), which has been adopted for the sake of clarity [*Blöschl*, 1991]. Changes over the day are assumed to be less important when simulating over several weeks or months. For very shallow packs the assumption of a homogeneous 25 × 25 m area is unrealistic [*O'Neill and Gray*, 1973]. To account for the effect of protruding boulders and bare patches, rather arbitrarily, albedo is not allowed to exceed a value of 0.7 at snow depths less than 10 cm.

Incoming long wave radiation is parameterized on the basis of air temperature and water vapor (an Ångström-type relation), as well as cloudiness. No correction is applied for the influence of surrounding terrain. Turbulent fluxes are parameterized by a wind function [*Blöschl et al.*, 1987]. The state of precipitation is discriminated on the basis of a fixed wet bulb temperature of 1°C [*Steinacker*, 1983].

Coupled heat and mass flow within the snowpack is simulated by a multilayer model following approaches of *Anderson* [1976] and *Colbeck and Davidson* [1973]. The state variables are dry density, liquid water content and snow temperature. The governing partial differential equations are approximated by finite differences and solved by an explicit scheme [*Siemer*, 1988]. This solution provides snow surface temperatures for estimating outgoing long wave radiation based on the Stefan-Boltzmann law. *Blöschl and Kirnbauer* [1991*a*] and *Blöschl* [1991] showed at the site scale that both the model of internal processes and the parameterizations of the energy balance components are applicable to conditions at Kühtai.

The distributed grid model presented here is computationally quite demanding. On a typical scientific workstation (e.g., RISC processor, 20 MIPS (megainstructions per second)) one run (2 months, 15,000 grid elements) would require about one day. More detailed information on the model and its computer time requirements is provided by *Blöschl and Kirnbauer* [1991*a*].

In this study, simulation runs are performed only for the

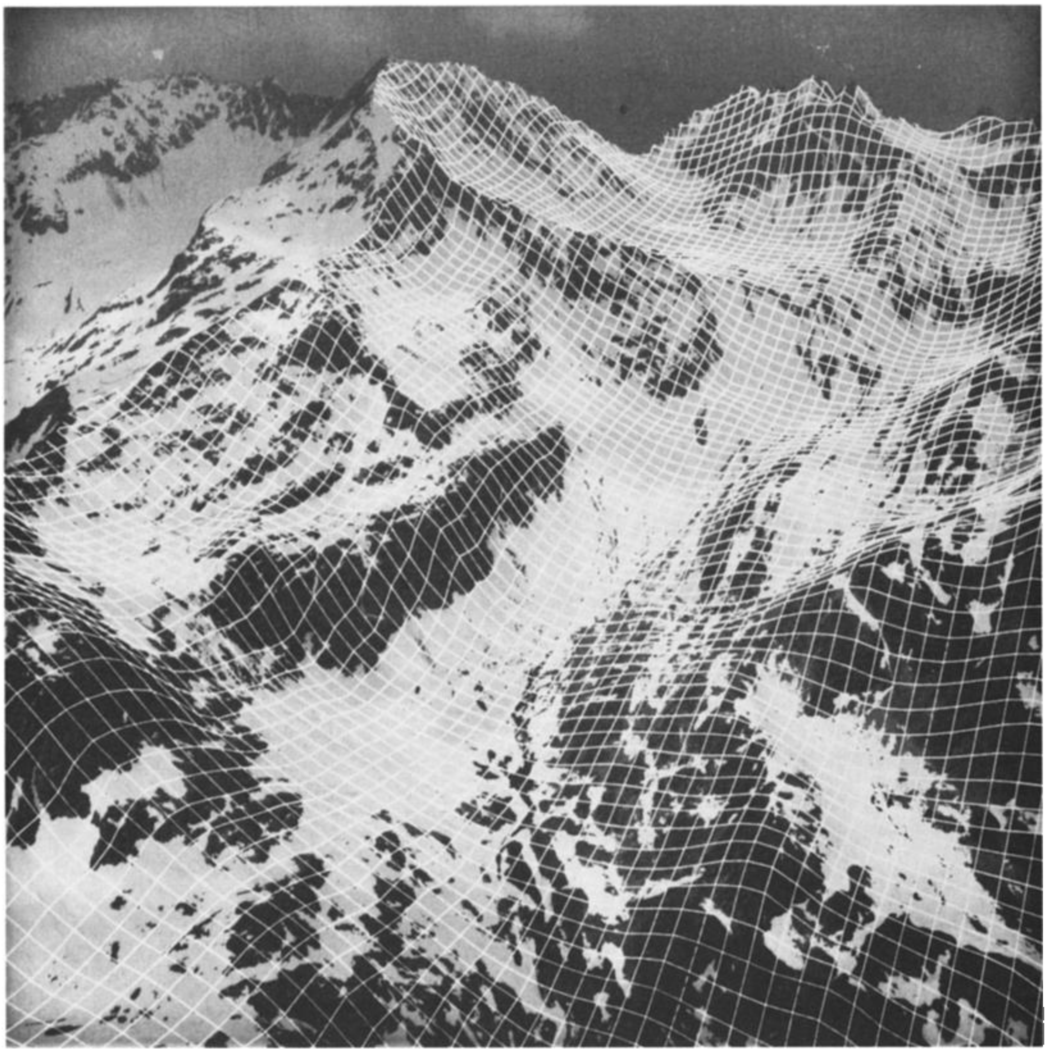

Fig. 4. Air photo of the upper part of the Längental catchment on June 26, 1989, showing grid elements 25 × 25 m. (By permission of Bundesministerium für Landesverteidigung.)

ablation period. Therefore, a good estimate of the spatial variation of snow cover parameters within the basin for late April 1989 is needed. Snow cover parameters include water equivalent and the thermal and hydraulic state of the pack. *Blöschl et al.* [1990] showed errors introduced to snowmelt simulations by an unknown thermal and hydraulic state of the basin snow cover to be insignificant as soon as 9 hours after model start once the pack has ripened. Therefore, zero heat and water storage is assumed as the initial state of the snow cover. However, the distribution of water equivalent is quite important.

The interpolation between measurements at "points," which in fact is the task, is difficult since water equivalent may vary greatly over small distances. *Elder et al.* [1989] recently presented an excellent review on this question and approached the interpolation problem by classifying a great number of depth measurements into terrain and radiation classes. Since in this study both logistical constraints and avalanche hazard disallowed the collection of many samples a more explicit interpolation scheme was adopted based on topographical features such as elevation, slope and local relief.

The relation of water equivalent to elevation is largely governed by climatic conditions [*McKay and Gray*, 1981; *Dickison and Daugharty*, 1984]. Here, a linear relation to elevation is assumed based on a best fit to the field data.

Due to the influence of wind and gravity, water equivalent tends to decrease as slope increases. However, there is a particularly large scatter about this relation [*Elder et al.*, 1989; G. R. Yates, personal communication, 1990]. *Golding* [1974] found a 6 to 8% decrease per 10% slope increase for an alpine catchment in Alberta. *Witmer* [1984] reports a linear decrease of snow depth between 35° and 50° with zero depths above that limit for the Swiss Alps. K. Elder (personal communication, 1990) found no snow on slopes steeper than 70° based on a 5 m grid in a steep Sierra Nevada catchment in California. G. R. Yates (personal communication, 1990) found this limit to be 45° based on a 90 m grid in a large basin in Colorado whereas in the Längental it is 60° on the basis of a 25 m grid [*Blöschl and Kirnbauer*, 1991*b*]. These findings indicate that the maximum slope is related to the spacing of the underlying grid. However, it is believed that the maximum slope of snow accumulation is more dependent on climate than on grid size. In California, it is basically a maritime climate, and snow is deposited at air temperatures close to 0°C. This means that the snow is warm

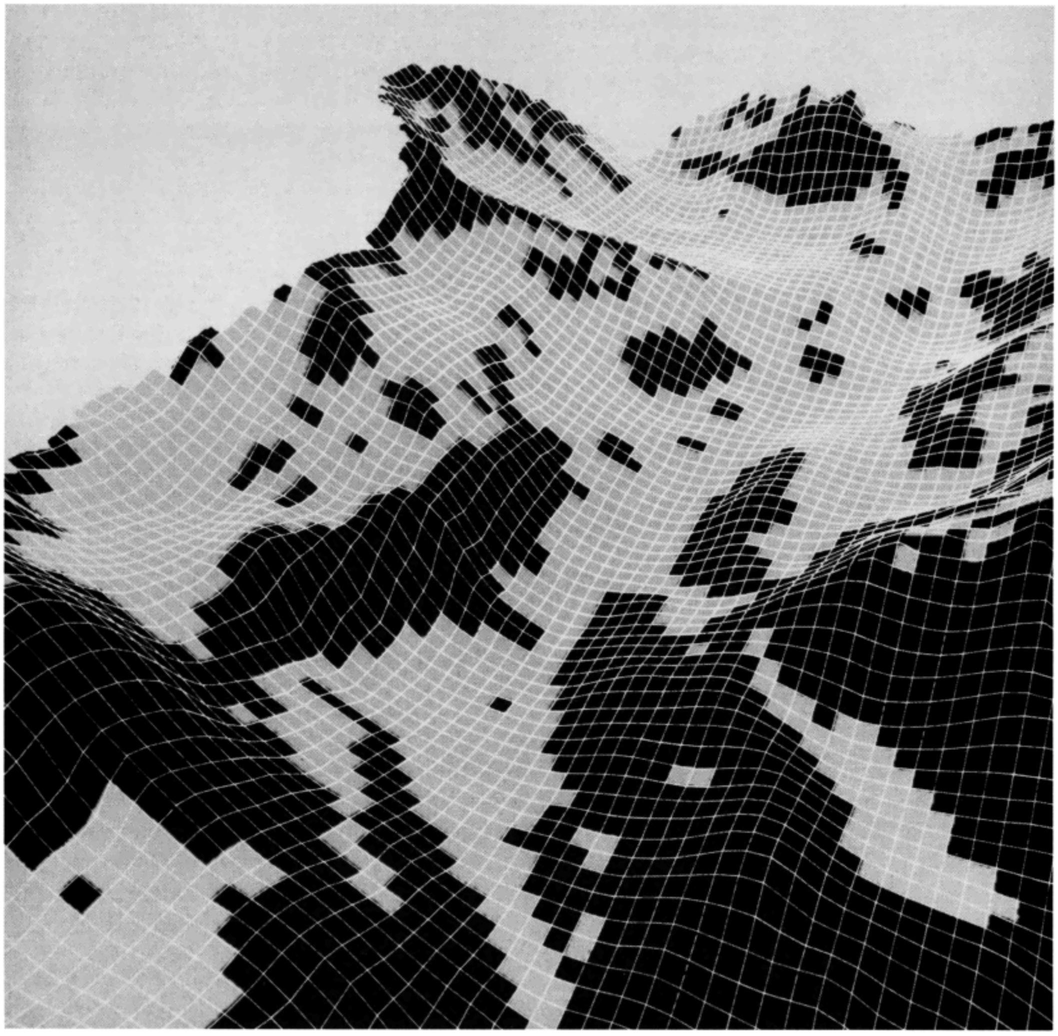

Fig. 5. Simulated snow cover on June 26, 1989. Dark areas denote bare ground and light areas denote snow cover.

and metamorphoses rapidly and the sintering allows the snow to stick to steeper surfaces and form a strong bond rapidly. Colorado, on the other hand, has a cold, continental climate and snow is deposited at low temperatures. The snow is very dry and metamorphoses very slowly. This means that it will not stick to steep slopes very well, and is inherently unstable. It is easily redistributed by wind and sloughs and avalanches easily from the steeper slopes as they load up. As to snow deposition, the climate of the Längental, Tirol is between those of California and Colorado and the choice of 60° as a critical slope appears to be justified. Based on the above considerations, water equivalent as a function of slope is assumed to be constant between 0° and 10° and to decrease linearly to zero between 10° and 60°.

The inclusion of the local relief primarily addresses the influence of wind drift. In a basin of hilly topography in the Canadian Arctic *Woo et al.* [1983*a*] found water equivalent to vary on average from 30% on hilltops to 300% in gullies as compared to flat areas. *Golding* [1974] reports mean water equivalents of 70% on ridge tops and 170% at valley bottoms for an Albertan basin. The more moderate influence of topography in Alberta as compared to the Arctic appears to be due to the forest cover. In this study, a relation to terrain curvature as derived from terrain data is arbitrarily chosen. Tops having a curvature of 0.02 m^{-1} (equivalent to 50 m radius) or greater are assumed to be snow free. Gullies with the same curvature are assigned 200% of the water equivalent of plane areas and linear interpolation is used in between.

The above assumptions give the following expression for interpolating water equivalent:

$$we = (a_1 + a_2 z)(1 - f(slope)(1 + a_3 curv) \qquad (1)$$

where we is the water equivalent (≥ 0), a_1 and a_2 are coefficients fitted to the field data, z is the elevation, $f(slope)$ is the influence of slope (degrees) (0 if $slope < 10°$, $slope/60$ otherwise), a_3 is set to 50 m and $curv$ is the terrain curvature (per meter) (>0 in gullies and <0 on tops).

Figure 2 shows measured water equivalent versus values predicted by the interpolation scheme. The relatively good fit of the data appears to derive from two facts: (1) sites were selected not randomly, but as being representative of a certain terrain type and (2) each data point represents the average over a 50 × 50 m area.

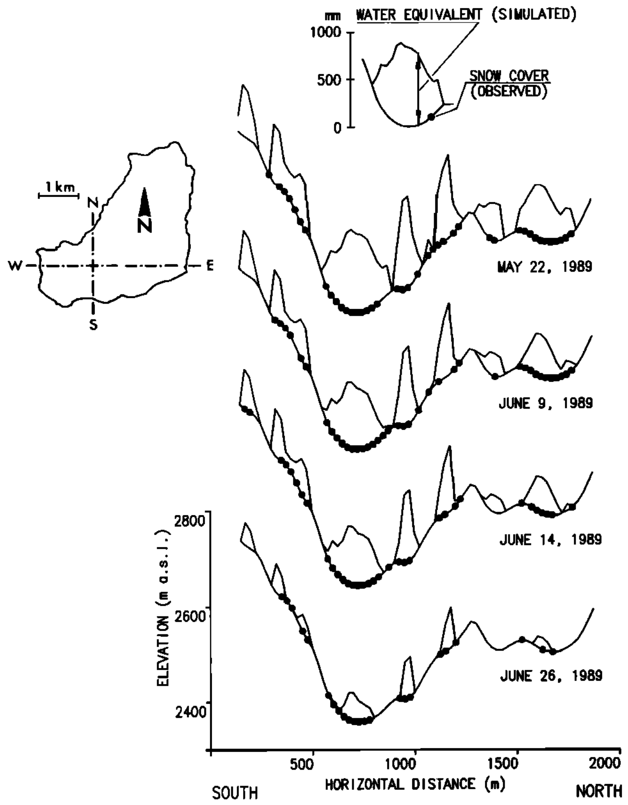

Fig. 6. Cross section showing topography, simulated water equivalent (line) and observed snow cover (solid circles) for south-north transect. Inset map shows location of the cross section in the Längental catchment.

4. Strategy of Model Evaluation

The distributed parameter model presented represents a number of complex snow accumulation and snowmelt processes and, therefore, consists of numerous model components. In fact, most of them can be questioned. Model evaluation based on an integrated value such as runoff is difficult because individual components cannot be disentangled from observations of their collective effect. Obviously, distributed data are more appropriate. As a first step, the focus was on snowmelt only, excluding runoff processes. This reduces the complexity of the system analyzed and yields a more precise evaluation of the snowmelt component. Clearly, a model evaluation is efficient when it allows for discriminating between alternative model assumptions. This is the case when the data tested are sensitive to these assumptions and the uncertainty introduced by other error sources is comparatively small. According to their sensitivity, different data may address different processes. This idea is adopted in this study.

Model performance is assessed on the basis of spatially distributed snow cover data. The relative importance of snowmelt processes varies within the basin and is, to some degree, related to topography. Accordingly, an analysis as a function of terrain features may be expected to identify individual processes. The following is a nonexhaustive attempt at relating snowmelt processes and phenomena to terrain features. A more detailed account is given by *Obled and Harder* [1979].

Factors related to elevation include increase of precipitation and water equivalent with elevation and processes controlled by air temperature such as turbulent heat exchange and the transition from rain to snowfall.

Factors related to slope include variations of water equivalent as induced by wind drift and avalanches and variation in direct beam solar radiation and solar radiation reflected by surrounding terrain.

Factors related to aspect include solar radiation (albedo) and variations of water equivalent as affected by prevailing wind directions.

Some phenomena cannot be reasonably related to a single terrain feature. These include variations of water equivalent at a regional scale; deposition affected by the microrelief; vegetation; avalanches; water equivalent on gullies and ridges as affected by wind drift; long wave terrain emission; and shading of solar radiation. Effects of these phenomena may be seen in map views, perspective views and cross sections of the catchment.

5. Results

Figure 3*a* shows the initial snow cover on April 24, 1989 as used for the model start. Figures 3*b* and 3*c* show the spatial distribution of simulated and observed snow cover for June 26, 1989. There is a good agreement of both snow-covered area and patterns. Simulated and observed snow-covered areas are 33 and 31%, respectively. However, snow cover is overestimated in the southeastern part of the basin and slightly underestimated in the northeastern part. Overall, the simulation exhibits fewer small patches of snow.

Figures 4 and 5 show an aerial photo of the upper part of the basin on June 26, 1989 along with the simulation results. Figure 4 indicates that there are marked subgrid variations in real snow cover. These have to be taken into account when assessing simulated patterns. There is a strong tendency toward more snow in gullies (Figure 5) which appears to be quite realistic. At the base of the steep cliff at the center of the photo there is a marked underestimation of snow cover.

South-north and west-east sections of the Längental catchment are presented in Figures 6 and 7, respectively, with observed snow cover and simulated water equivalent indicated. For a complete agreement water equivalent greater than zero should correspond to observed snow cover as marked by solid circles. Again, at the valley floor under steep slopes snow cover is underestimated (Figures 6 and 7). The variability in simulated water equivalent over the cross section appears to be large. When comparing sequential sections, errors in water equivalent may be detected to some degree. Particularly in the eastern part of the basin (Figure 7) water equivalent is overestimated.

Figure 8 shows an evaluation of simulation errors on an element-by-element basis for June 26. The elements are subdivided into classes according to slope and aspect separately for the upper and the lower part of the basin. The percentage denoted by "too late" refers to elements with snow cover simulated and bare ground observed, i.e., an overestimation of snow cover. As would be expected from the previous figures the agreement is good, which is reflected in generally low magnitudes of the error. In Figure 8 there is a certain symmetry about west and east facing slopes whereas the graph for north and south facing slopes is nearly antisymmetric.

Analogously to Figure 8, errors on June 26 are plotted versus elevation in Figure 9. To isolate elevation effects,

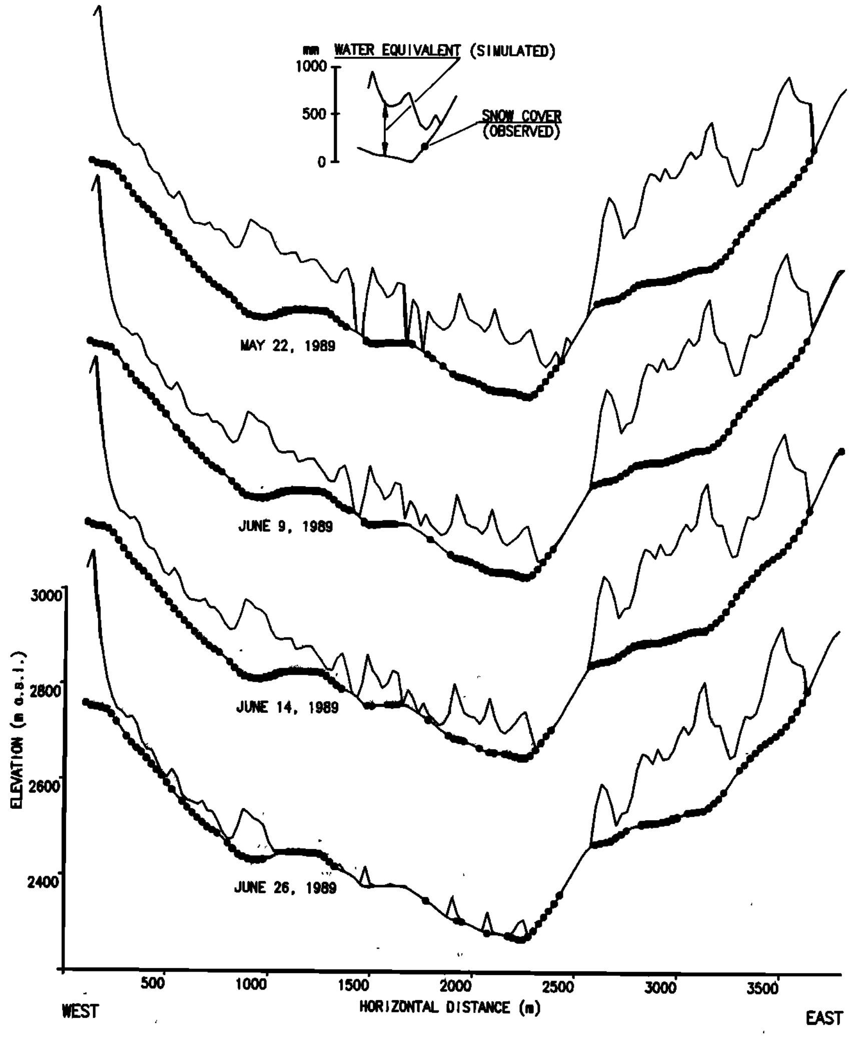

Fig. 7. Cross section showing topography, simulated water equivalent (line) and observed snow cover (solid circles) for west-east transect. For the location of the cross section see inset map in Figure 6.

only slopes less inclined than 20° are included. Below 2200 m above sea level there is a marked overestimation of snow cover on both north and south facing slopes whereas at higher elevations the agreement is good.

6. Discussion

Overall, the results outlined above show good agreement of observed and simulated snow cover patterns. However,

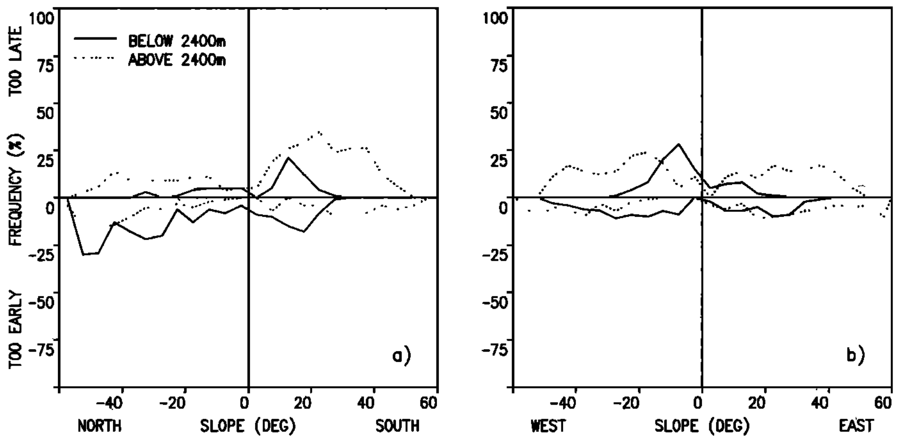

Fig. 8. Percent errors in snow cover for various slope and aspect classes, June 26, 1989 (too late: snow cover simulated, bare observed; too early: bare simulated, snow cover observed).

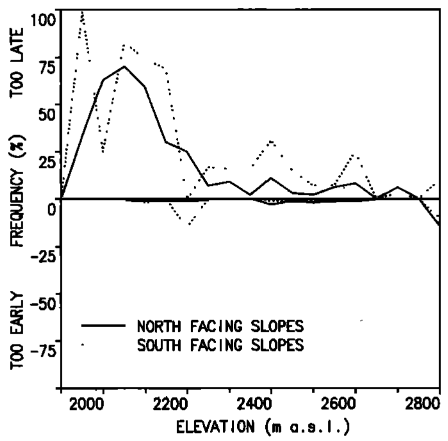

Fig. 9. Percent errors in snow cover versus elevation, June 9, 1989 (too late: snow cover simulated, bare observed; too early: bare simulated, snow cover observed).

simulated patterns tend to exhibit less dispersion of individual elements (Figures 3 to 5). This is attributable to subscale processes such as variations in snow deposition induced by the microrelief and microclimate. One way to model this type of scatter is by introducing a stochastic component for distributing water equivalent over the basin. Spatial correlation parameters as determined by *Elder and Dozier* [1990] could be used in such an analysis. Figures 4 to 7 also indicate an underestimation of snow cover at the base of steep slopes and cliffs which becomes more obvious as the season progresses. Clearly, these errors derive from redistribution processes such as avalanching, sloughing and wind drift.

Figures 3 and 7 indicate an overprediction of snow cover and water equivalent in the southeastern part of the basin which is formed by three prominent cirques. There are some indications that errors derive from long wave emission from the surrounding terrain which is not accounted for in the model. The configuration of surrounding ridges is such that rock walls may substantially enhance energy input through long wave emission, particularly when they are bare. The added energy may be equivalent to 500 mm melt over an entire snowmelt season [*Olyphant*, 1986]. On the basis of a comparison of sequential sections in Figure 7 it is speculated that in this study errors are of the order of 300 mm.

An underestimation of snow cover on north facing slopes and an overestimation on south facing slopes is found in Figure 8. This gives some indications that errors are related to solar radiation and specifically to albedo. There appear to be three reasons why this is so: (1) Albedo is well known to decrease with increasing grain size associated with metamorphism [*Colbeck*, 1988]. On south facing slopes, therefore, albedo may be expected to decrease more rapidly with time as more energy is available for metamorphism. (2) Albedo varies with changes in solar zenith angle [e.g., *Marshall and Warren*, 1987]. This is clearly an important factor in rugged alpine topography where albedo will be significantly lower on south facing slopes than on north facing slopes during nearly all times of day. In this analysis, however, albedo was assumed to be a function of the age of the snow surface only. (3) An overestimation of albedo along with an overestimation of turbulent or long wave input would produce the same results. This is because solar radiation is related to aspect whereas the latter fluxes are not. On north facing slopes errors in shortwave gain are smaller than on south facing slopes and, therefore, are more likely to be compensated by other error sources.

Figure 9 shows a marked step in the relationship of errors in snow cover to elevation. This may be easily traced back to the threshold temperature which discriminates rain from snow. Observations of the state of precipitation at the snow monitoring station on the preceding days (June 3 and 5) showed that rain at wet bulb temperatures slightly below zero was quite frequent. Given the observed temperatures the elevation of transition from rain to snow may be assumed to have varied around 2200 m above sea level, whereas the model predicted 1900 m with a threshold temperature of 1°C. This result illustrates that it is difficult to accurately predict the aggregational state of precipitation. One reason for the misclassification may be a seasonal trend in threshold temperatures with lower values in summer. Based on a study in the Swiss Alps *Rohrer* [1989] reports a tendency toward lower threshold temperatures in summer as compared to those in winter.

7. Conclusions

The simulation of snow cover patterns indicates that the basic model assumptions are realistic. Comparisons of simulated and observed depletion patterns are shown to be capable of addressing individual snow cover phenomena and model components. Model inadequacies may be attributed to two circumstances. (1) Processes not included in the model, such as redistribution of snow by avalanches, may play a role. Particularly, the effect of enhanced energy input to the snow cover in cirques by long wave terrain emission appears to be important. (2) Inaccurate values of model parameters may have been used; analysis of snow cover patterns suggests values for albedo and the transition temperature from rain to snow may be suspect. Results suggest that on south facing slopes albedo is significantly lower as compared to north facing slopes. The transition temperature appears to vary from storm to storm. The effect of inaccurate model parameters is dealt with in the companion paper.

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