Soil Vegetation-Atmosphere Transfer Schemes and Large-Scale Hydrological Models (Proceedings of a symposium held during the Sixth IAHS Scientific Assembly at Maastricht, The Netherlands, July 2001). IAHS Publ. no. 270, 2001.

Spatially-distributed snowmelt, water balance and streamflow modelling for a large mountainous catchment: Boise River, Idaho, USA

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Abstract As a demonstration of the potential of spatially-distributed hydrological modelling for large catchments in mountainous areas, a snow simulation model and a water balance/streamflow model have been linked together and applied to a 2150 km^2 portion of the Boise River catchment in Idaho, USA. The snow model simulates all of the energy fluxes into and out of the snowpack and requires spatial field time series of all of the major meteorological inputs. It is being run at 250 m grid resolution with a 3-h time step. The output spatial fields of this model are aggregated in space and time and are used as input to the water balance model. The water balance model simulates all of the major land surface processes using algorithms that are physically based but are constructed so as to be appropriate for this large spatial scale. It makes extensive use of geographic information systems for parameterization. The model operates on a daily time step and can have a variable cell size; for the Boise River, a 1 km^2 grid cell size is being used. These models are intended to address water and natural resource management issues, but the concepts of finding appropriate mathematical formulations and of matching the modelling scale with the process scale have broader significance.

Key words spatially-distributed hydrological modelling; snowmelt; energy balance; water balance; streamflow; geographic information systems; Boise River, Idaho, USA

INTRODUCTION

Hydrological modelling of large catchments in mountainous areas poses many challenges due to the great spatial variability of meteorological inputs and land surface characteristics and due to the issues involved in finding appropriate representations of physical processes at this scale. As a demonstration of the potential of spatially distributed hydrological modelling for large catchments in mountainous areas, an energy budget snow simulation model and a water balance/streamflow model have been linked together and applied to a portion of the Boise River catchment in the state of Idaho, USA. This modelling exercise is intended to show the necessary data processing techniques for, the feasibility of and the information that can be obtained from spatially distributed models and thus help to usher in a new generation of model for operational streamflow forecasting and water and natural resource management.

TEST AREA AND DATA SOURCES

The test area is a 2150 km^2 portion of the Boise River catchment, Idaho, USA. The elevation range is 1000–3200 m, with most of the area either sparse to moderately dense coniferous forest (55%) or shrubland (20%). Average annual precipitation ranges from about 500 mm in the lower elevations to over 1500 mm in the highest mountains, most of which occurs as snow. The catchment is a major part of a large system of reservoirs and canals providing water for irrigated agriculture.

There are eight meteorological stations in and near the catchment that are available to provide input data. Three of these stations have precipitation and temperature data at a 15-min resolution; five stations (part of the Natural Resources Conservation Service [NRCS] SNOTEL network) have precipitation, temperature, and snow water equivalent at a 3-h resolution; three of these five also have solar radiation, wind speed, relative humidity, and snow depth sensors.

A digital elevation model (DEM) is available with a 250-m resolution. A detailed digital vegetation map at a 30-m resolution is available, and this was aggregated to match the DEM. A soil texture map was derived from the NRCS State Soil Geographic (STATSGO) database.

SNOW SIMULATION MODEL

The model used to simulate the accumulation and melt of snow is called ISNOBAL (Marks *et al.*, 1999). It is a detailed energy budget model that has been successfully applied in several areas of the western USA. It is part of a modelling software package called Image Processing Workbench (IPW—see http://cirque.ars.pn.usbr.gov/~ipw for program descriptions and literature references).

The model is driven by spatial field inputs of net solar radiation, thermal radiation, temperature, precipitation, dew point temperature, wind speed and snow cover albedo to compute the energy and mass balance of the snow for each time step and at each grid cell in the area modelled. The model has been set up for the Boise River to run with 3-h time steps over a 250-m resolution grid.

An extensive set of data preparation procedures was developed to utilize the available meteorological data and process them into the form needed by the model. Net solar radiation fields required a multi-step procedure, involving: computation of topographically-corrected clear-sky incoming radiation using the IPW program TOPORAD; reduction of clear-sky radiation by a cloud cover factor estimated from measured radiation data; application of forest canopy corrections developed by Link & Marks (1999); and subtraction of reflected radiation due to snow albedo, computed by the IPW program IALBEDO. For thermal radiation, clear-sky fields were first computed using the IPW program TOPOTHERM, then they were multiplied by a cloud cover enhancement factor derived from measured data, and finally they were corrected for forest canopy effects, again using an algorithm from Link & Marks (1999). Precipitation and temperature spatial fields were interpolated from the data at the eight meteorological stations by an elevationally detrended kriging procedure (Garen *et al.*, 1994; Garen, 1995). Dew point temperature was computed from air temperature and relative humidity data at three stations, but these were not sufficient to do a spatial interpolation with the kriging procedure, so an alternative procedure was used. For this, the three values were averaged and considered to be appropriate at an elevation of 2100 m (the average elevation of the three stations). These values were compared to upper air soundings at the nearby airport in Boise, which are available twice a day. A dew point profile with elevation was developed by taking the surface dew point temperature from the upper air sounding (at 900 m) and the computed average dew point at 2100 m, but then above this, trending the dew point towards the upper air sounding value until it equalled this value at an elevation of 2800 m and above. This profile was then applied to the DEM to obtain spatial fields for each 3-h time step. The wind speed data available at three sites were averaged for each time step, and this same value was used for the lower and middle elevations, but it was increased up to a maximum of a factor of 1.5 in the highest elevations.

The energy and water fluxes simulated by the model are depicted in Fig. 1. The surface snow layer is where all of the energy exchanges with the atmosphere occur; these processes do not penetrate very far, so the thickness of this layer is set at a physically reasonable value of 0.25 m in the model. The lower layer is simply the remainder of the snow cover. Both layers are assumed to be homogeneous and are characterized by their average temperature, density, and liquid water content. The model assumes that energy is transferred between the surface layer and the lower layer and between the lower layer and the soil by conduction and diffusion. At each time step, the model computes the energy balance and the snow surface temperature and then adjusts the temperature and specific mass of each layer. If the computed energy

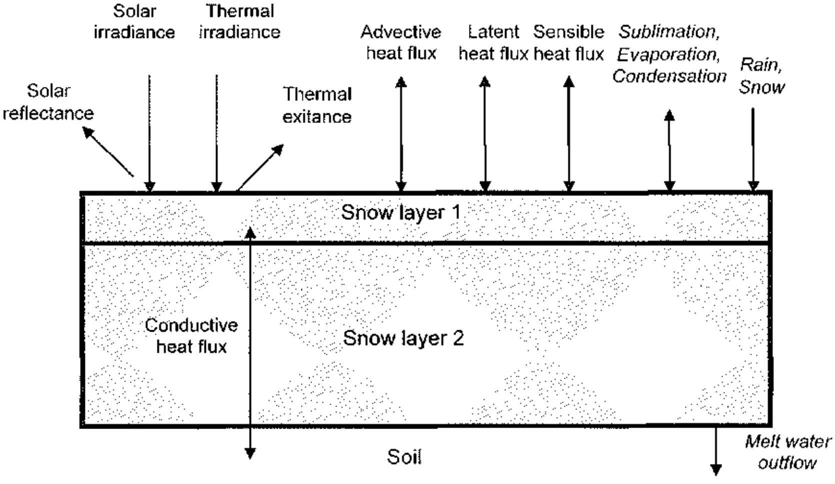

Fig. 1 Diagram of snow model components (energy fluxes in normal type, water fluxes in italics).

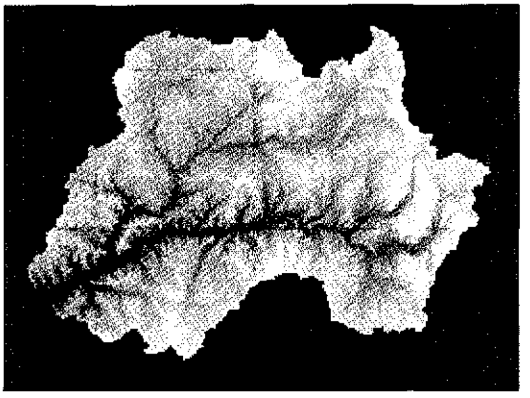

Fig. 2 Simulated snow water equivalent field for Boise River catchment, 1 April 1998; values range from zero (black) to 1100 mm (white).

balance is negative, the cold content, or the energy required to bring the temperature of the snow cover to 0°C, is increased, and the layer temperature decreases. If the energy balance is positive, the layer cold content is decreased until it is zero. Additional input of energy causes the model to predict melt. If melt occurs, it is assumed to displace air in the snow cover, causing densification and increasing the average liquid water content of both layers. Liquid water in excess of a specified threshold becomes meltwater outflow. Though meltwater is usually generated in the surface layer, outflow is removed from the lower layer. The thickness of the surface layer remains constant until the lower layer is completely melted. At that time, the model treats the snow cover as a single layer.

The model produces spatial field output at each time step for all of the energy and mass components shown in Fig. 1. An example of the simulated spatial field of snow water equivalent for 1 April 1998 is shown in Fig. 2. The output fields needed for input to the water balance model are snow water equivalent, snowmelt, and snow evaporation, which are aggregated from 3 h to daily fields and from 250 m to 1 km spatial resolution. Simulations are being carried out for the time period October 1996 onwards.

WATER BALANCE AND STREAMFLOW SIMULATION MODEL

The water balance and streamflow simulation model used has been developed primarily at the Ruhr University in Bochum, Germany, with assistance from the NRCS National Water and Climate Center in Portland, Oregon, USA. The land surface hydrological and streamflow generating process descriptions and parameters are intended to be comprehensive and appropriate for catchments approximately 100–10 000 km^2 in size and to be based as much as possible on spatial data that can be manipulated within a geographic information system (GIS). The intention has been to develop a spatially-distributed model that can be used for operational streamflow forecasting and water management that improves upon the empiricisms, spatial lumping, and many calibration parameters of the generation of conceptual models developed decades ago but that is still commonly used in practice. Previous descriptions of the model have been given by Schumann & Garen (1998) and Geyer *et al.* (2000). The key concepts are outlined below.

The model describes the spatially-distributed water balance of a catchment on a daily time step. To parameterize the hydrological process algorithms, spatial catchment characteristics of vegetation, soil texture classes, elevation and derived topographic data are used. These spatial characteristics are stored within a GIS and are used in conjunction with other soil and vegetation information to derive the model parameter values for each grid cell.

The spatial resolution of the model can be variable, depending on the heterogeneity or homogeneity of catchment characteristics. For the Boise River, the high degree of topographic variability precluded the identification of significant zones of homogeneity, so a square uniform grid of one kilometre width was chosen. The model assumes each spatial unit is large enough to contain a stream channel (thus avoiding the need to describe flow interactions among grid cells) but that it is small enough to be able to represent the spatial distribution of meteorological inputs.

The central component of the water balance model is the soil storage, which represents the depth of water that is held in the root zone. To obtain the maximum capacity of this soil storage, the soil porosity is multiplied by the root depth for each grid cell. Porosity is determined from the soil texture class according to the values given by Rawls *et al.* (1983). The root depth depends on the type of vegetation and the seasonal development of the roots. Values of root depth by month of the year for the various vegetation classes were taken from different literature sources (e.g. Disse, 1995). Soil and vegetation GIS layers have a finer spatial resolution than the 1 km^2 grid cell used for the hydrological model. Therefore, the existence of different combinations of soil and vegetation within each 1 km^2 grid cell can be determined. By overlaying the soil porosity and the root depth maps, multiplying the two within the GIS, and spatially aggregating, a cumulative distribution function of soil storage capacity is derived for each 1 km^2 grid cell. At this point, it is in the form of a step function, but then it is approximated by a linear function to reduce computational requirements in the model. The approximated distribution function varies seasonally, due to the variation in the root depth. It is this function that is used to establish the variable capacity of the soil moisture storage, as depicted in Fig. 3.

The soil storage distribution function is used to describe the temporally changing portion of saturated area within each grid cell. If, after the addition of the effective precipitation (i.e. after subtraction of interception losses) or snowmelt, the increase of the soil water content does not exceed the variable soil storage capacity at any point of the grid cell, no saturated areas exist, and no surface runoff is produced. If the actual water content of the soil storage exceeds the storage capacity of a certain part of the grid cell, this part is saturated (left-hand portion of Fig. 3). The addition of precipitation or snowmelt (the layer of water above the prior storage level in Fig. 3) then generates surface flow over this saturated portion of the grid cell and increases the amount of saturated area.

However, the impact of topography on the saturation of the soil also needs to be considered. The $\ln(a/\tan\beta)$ topographic index used in TOPMODEL, introduced by Beven & Kirkby (1979), is a widely-used characteristic to describe the relationship between the drainage area (a) and the slope ($\tan\beta$) at each point of a catchment. Under

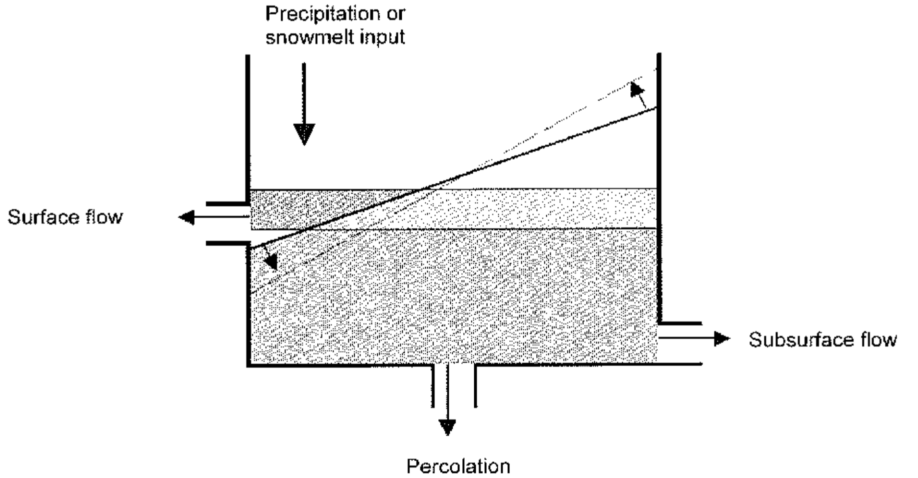

Fig. 3 Soil moisture storage showing the three outflows and how the distribution function of capacity and the topographic adjustment control the occurrence of surface runoff.

a given state of soil wetness within a catchment, the areas with a high value of the topographic index are more likely to be saturated than those with a lower value. In TOPMODEL, various simplifying assumptions are made to develop a quantitative relationship between the topographic index, catchment-wide soil moisture storage and saturated area. In the Bochum water balance model, it is not possible to use this relationship, as the same assumptions of spatial homogeneity used in TOPMODEL are not made. Instead, the index is used in a different way but still to consider the effect of topographic heterogeneity on saturation and runoff production.

To consider the influence of topography on the occurrence of saturated areas, the slope of the linear soil storage distribution function is adjusted. For each grid cell, the average of the topographic index $\ln(a/\tan\beta)$ is computed from the higher-resolution digital elevation data, and the mean value of the index for the catchment as a whole is also computed. It is expected that the grid cells with a topographic index above (below) the catchment average will be saturated earlier (later) and to a greater (lesser) extent than those with a lower (higher) value. This is considered by an increase or decrease of the slope of the distribution function of soil storage capacity. The amount of this increase or decrease is a linear function of the difference between the grid cell topographic index and the catchment average index, and it is regulated by a model parameter. The dashed line in Fig. 3 shows the case where the saturated part of a grid cell is increased as the slope of its distribution function is increased, corresponding to a grid cell with a topographic index above the catchment average. Increasing the slope therefore increases the dynamics of the soil water balance, as the amount of water that can be stored in the lower half of the range of soil storages is reduced, and the amount of direct runoff during rain periods is increased.

The soil storage controls not only the amount of surface runoff but also the subsurface runoff (lateral flow) and the percolation into the deeper soil, which is not affected by transpiration. Lateral flow and percolation are assumed to be gravity flow and are governed by Darcy's Law. The hydraulic conductivity is a nonlinear function of the saturated hydraulic conductivity and the relative fullness of the soil storage, as used by Brooks & Corey (1964). The storage of the percolated water and the base flow is described by two linear storages for the catchment, one for fast and one for slow base flow. Runoff is the sum of surface, subsurface and fast and slow base flow and is generated by each grid cell independently; it is routed to the catchment outlet based on its flow path distance as derived from the DEM.

Other model components parameterized with the help of GIS include the interception storage capacity and the actual evapotranspiration. The capacity of the interception storage is related to the leaf area index (*LAI*) using an approach suggested by Hoyningen-Huene (1993). The *LAI* can be estimated from remote sensing data or can be related to the type of vegetation and its seasonal development. Evaporation of intercepted precipitation is considered to occur at the potential evaporation rate. For estimation of evapotranspiration from the soil storage, a reference evapotranspiration for grassland is computed based on the well-known Penman-Monteith formula. To consider different types of vegetation and the steering effect of soil moisture, crop factors are used, which are taken from literature sources (e.g. Disse, 1995).

In the Boise River catchment, most of the streamflow originates as snowmelt, hence an accurate representation of these processes is crucial to simulating the water balance and streamflow. This was the motivation for using the detailed ISNOBAL model and using these results as input to the water balance model. It was felt that capturing the dynamics of the physical processes driving snowmelt required a greater level of detail and a greater time and spatial resolution than is required for the water balance, hence the higher resolution (3 h and 250 m) used in the snow model as opposed to that (1 day and 1 km) used in the water balance model.

The melt water outflow fields from the ISNOBAL runs are used as input to the soil system in the water balance model. This replaces precipitation input as long as snow exists on the grid cell. The water balance model then handles the soil moisture accounting and streamflow generation in the same manner as always.

MODEL VERIFICATION

Verifying a spatially distributed model is challenging because the reproduction of observed streamflow is no longer sufficient to determine whether the model components throughout the grid are functioning properly. As this work is ongoing, some approaches that are still under investigation will be mentioned here.

There are two possibilities for verifying the snow model. One is to compare observed snow water equivalent from measurement stations (NRCS SNOTEL sites in this case) with simulated values for the grid cell in which the station is located. Unfortunately, the two are not completely comparable, for three reasons: (a) the measurements are at a point, whereas the simulated values are an average over a grid cell; (b) the measurement sites are located on horizontal spots, even if the general topography is sloping, whereas in the model, the grid cells are characterized by this topographic slope; and (c) even if the general vegetation is forest, the measurement sites are in clearings, hence the effect of the trees on the energy balance of the measurement site is different from that for the grid cell, which is characterized as forested. The best way to determine if the snow model is functioning properly is to run a point version of the model using essentially the same inputs as for the spatial version, but making necessary adjustments to account for the differences noted above.

The other possibility for verifying the snow model is to compare simulated snow covered area with maps derived from satellite images. The difficulties with this are: (a) the resolution of the satellite images is usually much coarser than the model grid cell size; (b) the satellite images are intermittent in time so that a regular series of them is not available; (c) often part of the image is obscured by clouds, introducing some uncertainty into the identification of snow covered cells; and (d) snow covered area does not help determine if the amount of snow is correct and is therefore limited in its usefulness, particularly during winter when most of the catchment is snow covered anyway. Despite these difficulties, it is important to ensure that the model is realistically replicating the snow covered area, particularly during the active melt season.

Verifying the water balance model begins with reproducing observed streamflow at the catchment outlet, but another possibility for the Boise River is to compare soil moisture measurements available at a few of the SNOTEL sites in the catchment with the modelled soil moisture storage contents for the corresponding grid cells. As with snow water equivalent, there is a scale problem in comparing a point measurement with a simulated grid cell average, but it should be possible to obtain some useful information from this comparison. Other possibilities for verifying a spatial model, as mentioned by Geyer *et al.* (2000), such as special field campaigns to measure soil moisture along hillslope transects or conducting tracer experiments to evaluate the relative roles of surface/subsurface/groundwater flow, are not likely to occur in the Boise River catchment. Plans exist, however, to conduct these kinds of studies in catchments in Germany, which would still be useful in verifying the model structure and algorithms.

DISCUSSION AND CONCLUSION

This modelling study has had several goals:

(a) to develop a next-generation spatially-distributed hydrological model for operational use that includes a detailed simulation of snow accumulation and melt;
(b) to parameterize this model to take advantage of GIS and spatial data sets so as to increase the physical realism of the process algorithms and decrease the number of calibration parameters as compared to the generation of lumped conceptual models still in common use;
(c) to test the ability of collecting real-time solar radiation, wind, and humidity data in mountain locations, and demonstrate the use of these data in simulating snow accumulation and melt;
(d) to demonstrate the new types of hydrological information that can be extracted from a spatially-distributed model and how this spatial information could be used in water and natural resource management; and
(e) to demonstrate the ability of the spatial model to simulate streamflow, soil moisture, water fluxes and flow pathways within the catchment to improve streamflow prediction accuracy, particularly for unusual weather events or in cases of watershed change (e.g. fires, land-use change, climate change), and to improve the hydrological basis of erosion and water quality modelling.

In pursuing these goals, many modelling issues have had to be addressed. Much effort has gone into developing data processing procedures, both for the catchment characteristics and the meteorological input data, which have not always been straightforward. More fundamentally, a great deal of thought and judgment has gone into choosing the conceptualizations and parameterizations of the water balance model so that they are appropriate for the intended spatial scale while at the same time being faithful to represent the physical processes as realistically as possible given the spatial data and computing power available today. These same things must be considered when designing a model at other spatial and temporal scales, and at these scales, different choices will be made as to which processes to represent in a detailed fashion and which processes to represent more simply. The authors feel that the models described here represent a good balance of these considerations, and perhaps, by example, there are lessons here for model building at other spatial and temporal scales.

Acknowledgement The assistance of Josef Garen with graphics is greatly appreciated.

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