Stream flow simulation by WetSpa model in Hornad river basin, Slovakia

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ABSTRACT: Occurrence of severe floods in East European countries in the years 1997–2001 has drawn the attention of the public and also of the relevant authorities to the problem of flood protection. Combining advanced flood modeling with a GIS enables users and decision makers at various levels to investigate and assess proposed flood mitigation options and environmental impact assessments. In the Hornad catchment, which is one of the main tributary of the Tisza River, so far no rainfall-runoff forecasting model is in operational use. An application of a spatially distributed hydrologic model WetSpa working on a daily time scale is presented in this paper. The model combines elevation, soil and landuse data within GIS, and predicts flood hydrograph and the spatial distribution of hydrologic characteristics through a watershed. WetSpa model uses a modified rational method to calculate runoff and degree-day coefficient method to estimate the snow melt runoff based on temperature data. The runoff is routed through the basin along flow paths using a diffusive wave transfer model that enables to calculate response functions between any start and end point, depending upon slope, flow velocity and dissipation characteristic along the flow lines. The model is applied to the Hornad river basin located in Slovakia. The 4262 km² watershed is mountainous with elevations ranging from 171 to 1945 m.a.s.l and a mean slope of 17.6%. The mean annual precipitation in the catchment is about 664 mm. Daily hydrometeorological data from 1991 to 2000, including precipitation data from 44 stations, temperature data from 14 stations and evaporation data measured at 4 stations within and around the basin, are used as input data to the model. Three base maps, i.e. DEM, landuse and soil types are prepared in GIS form using 100 × 100 m cell size. Simulated hydrographs are compared to measured hydrographs which are available for the same 10-year period. Results of the simulations show a good agreement between calculated and measured hydrographs at the outlet of the basin and also in its main subbasins. The model predicts the daily hydrographs with a good accuracy, between 75 to 80 % according to the Nash-Sutcliff criteria. Since the spatial distribution of hydrologic characteristics can be obtained from the model outputs at each time step, the model has a great potential to analyze the effects of topography, soil type, land cover and the effects of landuse changes on the hydrologic behavior of a river basin.

1 INTRODUCTION

Rainfall-runoff computation by distributed hydrologic models and using GIS techniques has become increasingly possible, practical and popular. The models are becoming more capable for predicting flood and decision making in watershed management.

Grid-based GIS appears to be a very suitable tool for spatially distributed hydrologic modeling. Starting from a digital elevation model, hydrologic features of the terrain can be determined using standard GIS functions that operate on raster terrain data. Also, estimation of surface and soil related parameters becomes feasible by combining soil type and land use data in raster format. Flow routing can be achieved by tracking the water throughout the cell network along topographic flow paths.

The WetSpa model used in this study is a simple grid-based distributed runoff and water balance simulation model that runs on an hourly or daily time step. It predicts hourly/daily overland flow occurring at any point in a watershed, hydrograph at the outlet, and provides spatially distributed hydrologic characteristics in the basin, in which all hydrologic processes are simulated within a GIS framework. Inputs to the model include digital elevation data, soil type, land use data, precipitation and potential evaporation time series. Stream discharge data is optional for model calibration.
In this paper, an application of the WetSpa model on a large catchment located in Slovakia on a daily time scale is presented.

2 WETSPA MODEL

The WetSpa model was originally developed by Wang et al. (1996) and adapted for flood prediction by De Smedt et al. (2000) and Liu et al. (2003). For each grid cell, four layers are considered in the vertical direction as vegetation zone, root zone, transmission zone and saturated zone. The hydrological processes considered in the model are precipitation, interception, depression, surface runoff, infiltration, evapotranspiration, percolation, interflow, ground water flow, and water balance in the root zone and the saturated zone. The total water balance for a raster cell is composed of the water balance for the vegetated, bare-soil, open water and impervious parts of each cell. This allows one to account for the non-uniformity of the land use per cell, which is dependent on the resolution of the grid. The processes in each grid cell are set in a cascading way, which means that an order of occurrence of the processes is assumed after a precipitation event. A mixture of physical and empirical relationships is used to describe the hydrological processes in the model. The model predicts peak discharges and hydrographs, which can be defined for any numbers and locations in the channel network, and can simulate the spatial distribution of catchment hydrological characteristics.

Incident rainfall first encounters the plant canopy, which intercepts all or part of the rainfall until the interception storage capacity is reached. The excess water reaches the soil surface and can infiltrate the soil zone. Water that does not infiltrate goes to depression storage, and is diverted as surface runoff along with depression storage simultaneously. The depression storage is subject to evaporation and further infiltration. The sum of the interception and depression storage forms the initial losses at the beginning of a storm, and does not contribute to the storm flow. A fraction of the infiltrated water percolates to the groundwater storage and some is diverted as interflow. Soil water is also subject to evapotranspiration depending on potential evapotranspiration rate and available soil moisture. Groundwater discharges to the nearest channel according to the groundwater storage and the recession coefficient. Evapotranspiration from groundwater storage is also accounted, which gives the effect of a steeper recession during dry period. Total runoff from a grid cell is then the summation of surface runoff, interflow and groundwater discharge.

For each grid cell, the root zone water balance is modeled continuously by equating inputs and outputs:

\[ D \frac{\partial \theta}{\partial t} = P - I - V - E - R - F \]  

(1)

where \( D \) [L] is the root depth, \( \Delta \theta \) [L3L-3] is the change in soil moisture, \( \Delta t \) [T] is the time interval, \( I \) [L-1] is the initial abstraction including interception and depression losses within time step, \( V \) [LT-1] is the surface runoff or rainfall excess, \( E \) [LT-1] is the actual evapotranspiration from the soil, \( R \) [LT-1] is the percolation out of the root zone, and \( F \) [LT-1] is the amount of interflow in depth over time. The rainfall excess is calculated using a moisture-related modified rational method with potential runoff coefficient depending on the land cover, soil type, slope, the magnitude of rainfall, and the antecedent soil moisture. The values of potential runoff coefficient are taken from literature and a lookup table is generated, linking values to slope, soil type and land use classes. Evapotranspiration from soil and vegetation is calculated based on the relationship developed by Thornthwaite and Mather (1955) as a function of potential evapotranspiration, vegetation type, stage of growth and soil moisture content. For the surface layer, actual evapotranspiration is computed as the area-weighted mean of the land use percentage, for which transpiration happens from the vegetated parts, evaporation happens from the bare soil, and there is no evaporation on impervious areas. A portion of the remaining potential evapotranspiration after the abstraction from soil and land surface is transpired from the groundwater water as a function of the groundwater storage. Finally, the total evapotranspiration is calculated as the sum of evaporation from interception storage, depression storage, and the evapotranspiration from soil and groundwater storage. Percolation and interflow are assumed to be gravity driven. The percolation out of the root zone is equated as the hydraulic conductivity corresponding to the moisture content as a function of the soil pore size distribution index (Eagleson 1978). Interflow is assumed to occur in the root zone after percolation and becomes significant only when the soil moisture is higher than field capacity. Darcy’s law and a kinematic wave approximation are used to estimate the amount of interflow generated from each cell, in function of hydraulic conductivity, the moisture content, slope angle, and the root depth.

The routing of overland flow and channel flow is implemented by the method of the diffusive wave approximation of the St. Venant equation:

\[ \frac{\partial Q}{\partial t} + \frac{\partial Q}{\partial x} = -c \frac{\partial Q}{\partial x} - \frac{\partial Q}{\partial t} \]  

(2)

where \( Q \) [L3T-1] is the discharge at time \( t \) and location \( x \), \( t \) [T] is the time, \( x \) [L] is the distance along the flow direction, \( c \) [LT-1] is the location dependent kinematic wave celerity and is interpreted as the velocity by which a disturbance travels along the flow path,
and $d [L^2 T^{-1}]$ is the location dependent dispersion coefficient, which measures the tendency of the disturbance to disperse longitudinally as it travels downstream. Assuming that the bottom slope remains constant and the hydraulic radius approaches the average flow depth for overland flow and watercourses, $c$ and $d$ can be estimated by $c = (5/3)v$, and $d = (vH)/(2S_0)$ (Henderson 1966), where $v [L/T]$ is the flow velocity calculated by the Manning equation, and $H [L]$ is the hydraulic radius or average flow depth. A linear approximate solution to the diffusive wave equation in the form of a first passage time distribution is applied (Liu et al. 2003), relating the discharge at the end of a flow path to the available runoff at the start of the flow.

$$U(t) = \frac{1}{\sigma \sqrt{2\pi t_{0}}} \exp\left[-\frac{(t - t_{0})^2}{2\sigma^2 t_{0}}\right]$$

(3)

Where $U(t) [T^{-1}]$ is the flow path unit response function, serving as an instantaneous unit hydrograph (IUH) of the flow path that makes it possible to route water surplus from any grid cell to the basin outlet or any downstream convergent point, $t_0 [T]$ is the flow time, and $\sigma [T]$ is the standard deviation of the flow time. The parameters $t_0$ and $\sigma$ are spatially distributed, and can be obtained by integration along the topographic determined flow paths as a function of flow celerity and dispersion coefficient.

$$t_0 = \int \frac{1}{c} dx$$

(4)

and

$$\sigma = \sqrt{\int \frac{2d}{c} dx}$$

(5)

Because, groundwater movement is much slower than the movement of water in the surface and near surface water system, and little is known about the bedrock, groundwater flow is simplified as a lumped linear reservoir on small GIS derived subcatchment scale. Considering the river damping effect for all flow components, overland flow and interflow are routed firstly from each grid cell to the main channel, and joined with groundwater flow at the subcatchment outlet. Then the total hydrograph is routed to the basin outlet by the channel response function derived from Equation 3. The total discharge is the sum of overland flow, interflow and groundwater flow, and is obtained by a convolution integral of the flow responses from all grid cells. One advantage of this approach is that it allows the spatially distributed runoff and hydrological parameters of the basin to be used as inputs to the model. Inputs to the model include digital elevation data, soil type, land use data, and measured climatological data. Stream discharge data is optional for model calibration. All hydrological processes are simulated within a GIS framework.

Because a large part of the annual precipitation is in the form of snow, a model based on daily temperature data is used to simulate snow melt. The conceptual temperature index or degree-day method is used in this study, because it is simple but has a strong physical foundation. The method replaces the full energy balance with a term linked to air temperature. It is physically sound in the absence of shortwave radiation when much of the energy supplied to the snowpack is atmospheric long wave radiation. The equation can be expressed as:

$$M = \text{Max}[0, C_{\text{snow}}(T - T_0) + C_{\text{rain}}P(T - T_0)]$$

(6)

Where $M [LT^{-1}]$ is the daily snowmelt, $T [°C]$ is the daily mean temperature, $T_0 [°C]$ is a threshold melt temperature, $C_{\text{snow}}$ is a melt-rate factor [L°C$^{-1}$T$^{-1}$], and $C_{\text{rain}}$ is a degree-day coefficient taking into account the heat contribution from rainfall [L$^{-1}$C$^{-1}$T$^{-1}$]. The critical melt temperature, $T_{0}$, is often intuitively set to 0°C. The melt-rate factor, $C_{\text{snow}}$, is an effective parameter and may vary with location and snow characteristics. However, in this study this parameter is assumed as a constant for model simplicity.

3 APPLICATION

3.1 Study area

The Hornad basin is located in Slovakia. The watershed has an area of 4262 km$^2$ up to Zdana station. A multi purpose reservoir called Ruzin is located in the centre of the basin (Fig. 1). The Hornad basin is a large mountainous catchment, with elevation ranging from 171 to 1945 m.a.s.l. The mean elevation of the catchment is 580 m; the mean slope of the catchment is about 17.6%. Some of the physiographical and hydrological characteristics of the Hornad basin are given in Table 1. Figure 1 shows the Hornad basin, topography, flow stations, and the Ruzin reservoir.

3.2 Basic grid maps

The DEM for the river basin was obtained from the Slovak Hydrometeorological Institute (SHMU), and converted to a 100 m grid size DEM, from which the drainage system and area were determined (Fig. 1). Land cover data were obtained from the third hierarchy CORINE geographic information system coverage from the European PHARE Project.
The final landuse map (Fig. 2) for this study has 100 x 100 m cell size and is composed of 6 different types of land cover. Based on this map 51.8% of the basin is covered by forest (17.8% coniferous forest and 34% mixed & deciduous forest). Grassland and pasture cover 22.7% of the area; 22.5% of the basin is covered by agriculture areas, and 2.9% is urban area. Water surfaces, which are mainly reservoirs within the basin, cover about 0.1% of the watershed.

Considering to the soil map (Fig. 3) there are 10 different soil textures in the catchment. The dominant soil texture is silt loam, which covers about 43% of the basin area; loam and sandy loam cover the area with 27 and 17% respectively.

3.3 Climatology

The basin has a northern temperate climate with four distinct seasons. January is the coldest month and July is the warmest month of the year. The highest amount of precipitation occurs in the period from May to August while in January and February there is usually only snow. The mean annual precipitation of the watershed based on 10 years data of 36 stations within the basin is 664 mm. It ranges from about 559 mm in Vysny Caj (230 m.a.s.l) to more than 1200 mm in the vicinity of the water divide from the analysis of Slovak Hydrometeorological Institute in Bratislava. The mean temperature of the catchment based on a 40-year period isothermal map is about 6.5°C. The annual potential evapotranspiration based on 10 years data of 4 stations inside and surrounding the basin is about 568 mm.

3.4 Input data

For this study, precipitation, temperature and discharge data were obtained from SHMU, whereas the potential evapotranspiration (PET) data were obtained from the Water Research Institute of Slovakia. The sets include daily precipitation for 44 stations, temperature for 14 stations, PET at 4 stations, and daily discharge data at 24 gauging stations. All these data are available for a 10-year period from 1991 to 2000. Flow stations containing daily discharge data at 24 locations are available inside the catchment, but only the Hornad outlet station, Zdana and the outlet stations of the main sub-basins are used for model calibration.

3.5 Model simulation

Once the required data are collected and processed for use in the WetSpa modeling platform, identification of spatial model parameters is undertaken. Terrain features at each grid cell including elevation, flow direction, flow accumulation, stream network, stream link, stream order, slope, and hydraulic radius are first extracted from the DEM. The threshold for delineating...
the stream network is set to 10, i.e. a cell is considered being drained by streams when the upstream drained area becomes greater than 0.1 km². The threshold value for determining subcatchments is set to 1000, by which 223 subcatchments are identified with an average subcatchment area of 19.1 km². When creating the grid of surface slope, a threshold value of minimum slope of 0.01% is considered; if the calculated slope is less than this threshold value the slope is set to 0.01% in order to avoid stagnant water or extreme low velocities (Fig. 4). The grid of hydraulic radius is generated with an exceeding frequency of 0.5 (2-year return period), for which the network constant and the geometry scaling exponent are set to 0.05 and 0.48, resulting in an average hydraulic radius of 0.005 m for the upland cells and 2.76 m at the outlet of the main channel. Next, the grids of soil hydraulic conductivity, porosity, field capacity, residual moisture, pore size distribution index, and plant wilting point are reclassified based on the soil texture grid by means of an attribute lookup table. Similarly, the grids of root depth, interception storage capacity, and Manning’s roughness coefficient, n, are reclassified from the land use grid, in which the Manning’s n for channels is linearly interpolated based on the stream order grid with 0.060 m^{-1/3}s for the lowest order and 0.030 m^{-1/3}s for the highest order. The grids of potential runoff coefficient and depression storage capacity are obtained by means of attribute tables combining the grids of elevation, soil and land use, for which the percentage of impervious area within an urban cell is set to 30%. As it can be seen in Figure 5, the non-forested and steeper areas generate a very high runoff coefficient, whereas the forested and gentle slopes generate...
less surface runoff. The calculated average potential runoff coefficient is 0.42 for the entire catchment.

The grids for precipitation, temperature and PET are created based on the geographical coordinates of each measuring station and the catchment boundary using the Thiessen polygon extension of the ArcView Spatial Analyst.

Finally, the grids of flow velocity, travel time to the basin outlet, as well as the standard deviation are generated, which enables to calculate the IUH from each grid cell to the basin outlet. Figure 6 shows the estimated average flow time from each grid cell to the basin outlet. Flow time for the most remote area is around 71 hours. The mean travel time for the entire catchment is 33 hours.

3.6 Results

The 10 years (1991–2000) measured daily precipitation, temperature, PET, and discharge data are used for model calibration. The calibration process is mainly performed for the global model parameters, whereas the spatial model parameters are kept as they are. The initial global model parameters are specifically chosen according to the basin characteristics as discussed in the documentation and user manual of the model (Liu & De Smelt 2004). The simulation results are then compared to the observed hydrograph at Zdana both graphically and statistically. The parameters of base temperature and degree-day coefficients are adjusted independently in order to get a proper fit of snowmelt, occurring normally in late February or early March. The initial groundwater flow recession coefficient is estimated by analyzing the baseflow, which is separated from the observed hydrograph. Adjustment of this parameter is necessary in accordance with the fitting of baseflow and the total flow volume. The interflow scaling factor is adjusted for the peak and recession part of the flood hydrograph, which is sensitive for both high and low flows. The
additional two parameters controlling the amount of surface runoff, i.e. the surface runoff exponent for a near zero rainfall intensity and the rainfall intensity corresponding to a surface runoff exponent of 1, are adjusted mainly for small storms, for which the actual runoff coefficients are small due to the low rainfall intensity. The initial soil moisture and active groundwater storage are adjusted by comparison of the hydrographs and water balance for the initial phase. The maximum active groundwater storage controls the amount of vapor transpired from the groundwater, and therefore can be adjusted by comparison of the flow volume during dry periods.

Figures 7 and 8 give a graphical comparison between observed and calculated daily flow at Zdana for the years 1991 and 1997. Figures 7 and 8 show that both the spring and summer flood hydrographs are well reproduced by the model. The simulation of snowmelt flood is important in this study as it not only contributes to the results of model evaluation, but also provides a reliable soil moisture estimation at the end of the snow melting period, which affects following rainfall runoff processes. The calibrated base temperature and degree-day coefficient are 0°C and 2.7 mm/d°C, whereas the heat contribution from rainfall to the snowmelt is not important for this case study. The calibrated groundwater flow recession coefficient at Zdana is 0.0155 d⁻¹, which is smaller than the calculated value, and gives a good estimation for the whole simulation period. The peak discharges, concentration time, and flow volumes are especially well predicted for the three summer floods of 1997.

Similar simulation results are obtained for other hydrological years. The model performance is satisfactory from the statistical evaluation results, in which the flow volume is 2.7% over-estimated, the Nash-Sutcliffe efficiency is 0.773, and the modified Nash-Sutcliffe efficiency is 0.763 and 0.814 respectively for high and low flows. This indicates that the model is able to consider the precipitation, antecedent moisture and runoff-generating processes in a spatially realistic manner based on topography, land use and soil type, resulting in a fairly high accuracy for both high and low flows, and the general hydrological trends being well captured by the model. The model outputs also show that 8.43% of the precipitation is intercepted by the plant canopy, 84.25% infiltrates to
the soil, 70.31% evaporates to the atmosphere, 20.70% recharges to the groundwater reservoir, and 29.41% becomes runoff, of which direct flow, interflow, and groundwater flow contribute 22.7%, 10.0% and 67.3% respectively. These values are reasonable in view of the catchment hydrological characteristics.

4 DISCUSSION AND CONCLUSION

In this paper an attempt is made to outline a method for estimating flood runoff in the Hornad basin by using detailed basin characteristics together with meteorological data as an input to the WetSpa spatially distributed model. To avoid the complexity inherent in estimating surface runoff, a simple but effective approach is presented where the whole basin is divided into grid cells, giving the possibility to simulate the hydrologic processes at reasonably small scale.

The generation of runoff depends upon rainfall intensity and soil moisture and is calculated as the net precipitation times a runoff coefficient, which depends upon slope, land use and soil type. Overland flow is routed through the basin along flow paths determined by the topography using a diffusive wave transfer model, while interflow and groundwater recharge are simulated using Darcy’s law and the kinematic approximation. Model parameters based on surface slope, land use and their combinations are collected from literature, which can be prepared easily using standard GIS techniques.

The model is tested on the Hornad mountainous catchment in Slovakia with 10 years of observed daily rainfall and evaporation data. Good agreement with the measured hydrograph is achieved.

As the model is aimed to predict floods and also because of lack of information about the reservoir operation condition during the 10-year period, the effect of the reservoir was not taken into account. The consequences of ignoring the operation of the reservoir can be seen in the dry period when the reservoir is operated for storing water. Hence, the results of 10 years simulation show a lesser accuracy for the low flows.

Since the spatial distribution of hydrologic characteristics can be obtained from the model outputs at each time step, the model is especially useful to analyze the effects of topography, soil type, and land-use on the hydrologic behavior of a river basin. Improvements of the interflow redistribution and combination with a physically based distributed groundwater model in order to evaluate the effect of groundwater table to the generation of surface runoff are the next steps in this study.

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