Application of WetSpa model for assessing land use impacts on floods in the Margecany–Hornad watershed, Slovakia


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Abstract

The spatially distributed hydrologic model WetSpa combines elevation, soil and land use data within GIS, to predict flood hydrographs and spatial distribution of hydrologic characteristics in a watershed. The model is applied to the Margecany–Hornad river basin (1,131 km²) in Slovakia. Daily hydrometeorological data from 1991–2000, including precipitation data from nine stations, temperature data from four stations and evaporation data measured at one station are used as input to the model. Three base maps, i.e. DEM, land use and soil type are prepared in GIS form, using 100 × 100 m cell size. Results of the simulations show good agreement between calculated and measured hydrographs. The model predicts the daily/hourly hydrographs with 75–80% accuracy according to the Nash–Sutcliffe criteria. For assessing the impact of land use changes on floods, the calibrated model is applied for a reforestation scenario, which considers a 50% increase of forest areas. The model results show that the reforestation scenario decreases the peak discharge by 12%. Investigation of peak discharges from the whole simulation period, shows that the scenario results are reduced by 18% on average, while for small discharges the reduction is even about 34%. The time to peak of the simulated hydrograph of the reforestation scenario is 20 hours longer than for the present land use.

Keywords

Flood prediction; hydrological modelling; land use impacts; WetSpa model

Introduction

The hydrologic effects of land use changes have been thoroughly described by Calder (1993), Ward and Robinson (1990) and summarized in De Roo et al. (2003). Major changes in land use that affect hydrology are deforestation, intensification of agriculture, drainage of wetlands, and urbanization. The most obvious influence of land use on the water balance of a catchment is on the evapotranspiration process (Calder, 1993). Different land use types have different evapotranspiration rates, because different crops have different vegetation cover, leaf area indices, root depths and albedo. Also, interception rates are different although the influence of interception is noticeable only during small storms (Ward and Robinson, 1990; Calder, 1993). But land use influences especially infiltration and soil water redistribution. An extreme example is the influence of build up areas on overland flow. Finally, land use and land management influence surface roughness, which controls overland and floodplain flow rates (De Roo et al., 2003).

Computer simulation modelling has been used for at least 35 years to study the effects of land use changes within catchments (see, e.g. Onstad and Jamieson, 1970; Hillman and Verschuren, 1988; Bulot et al., 1990; Ewen and Parkin, 1996; Niehoff et al., 2002).
Rainfall–runoff generation by distributed hydrological models and GIS techniques has become increasingly possible, practical and popular. The models are becoming more capable for predicting flood, land use impacts on floods, and decision making in watershed management. 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 grid-based distributed runoff and water balance simulation model that runs on an hourly or daily time step. It predicts hourly or daily overland flow occurring at any point in a watershed, and provides spatially distributed hydrologic characteristics in the basin. Inputs to the model include digital elevation data, soil type, land use, precipitation and potential evaporation time series. In this paper, an application of the WetSpa model is presented for a rather large catchment located in Slovakia for assessing the impacts of land use changes on floods.

WetSpa model

The WetSpa model was originally developed by Wang et al. (1997) 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: a canopy layer, the root zone, a transmission zone and the groundwater reservoir. The hydrological processes considered in the model are precipitation, interception, depression storage, surface runoff, infiltration, evapotranspiration, percolation, interflow, groundwater flow, and water balance in each layer. The total water balance for each raster cell is composed of a separate water balance for the vegetated, bare-soil, open water and impervious part of each cell. This allows us to account for the non-uniformity of the land use per cell, which is dependent on the resolution of the grid. For each grid cell, the root zone water balance is modeled continuously by equating inputs and outputs:

$$D \Delta \theta \Delta t = P - I - V - E - R - F$$

where $D$ [L] is the root depth, $\Delta \theta$ [L$^3$L$^{-3}$] is the change in soil moisture, $\Delta t$ [T] is the time interval, $I$ [LT$^{-1}$] is the initial abstraction including interception and depression losses within time step $\Delta t$, $V$ [LT$^{-1}$] is the rate of 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 coefficients depending on land cover, soil type, slope, the magnitude of rainfall, and the antecedent soil moisture.

For the surface layer, actual evapotranspiration is computed as an area-weighted mean of the land use percentage. Also a portion is transpirated from the groundwater water as a function of the groundwater storage. Percolation and interflow are assumed to be gravity driven. The percolation out of the root zone is equated as the hydraulic conductivity as a function of the soil moisture. Interflow is assumed to occur in the root zone after percolation 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:

$$\frac{\partial Q}{\partial t} = d \frac{\partial^2 Q}{\partial x^2} - c \frac{\partial Q}{\partial x}$$
where $Q$ [$L^2 T^{-1}$] 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 expresses the tendency of the disturbance to disperse longitudinally as it travels downstream. Assuming that 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 $d = (vH)/(2S_0)$, where $v$ [LT$^{-1}$] 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, relating the discharge at the end of a flow path to the available runoff at the start of the flow:

$$U(t) = \frac{1}{\sigma \sqrt{2\pi t H_0^2}} \exp \left[ -\frac{(t - t_0)^2}{2\sigma^2 H_0^2} \right]$$

(3)

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

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

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

(4)

Because groundwater movement is much slower than the movement of surface water, groundwater flow is simplified as a lumped linear reservoir for each subcatchment. 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. Since, a large part of the annual precipitation is in the form of snow, the conceptual temperature index or degree-day method is used to simulate snow melt.

**Application**

**Study area**

The Hornad river located in Slovakia, has a drainage area of 1,131 km$^2$ up to the Margecany station (in this paper simply named the Margecany catchment) located just above a multi purpose reservoir called Ruzín. Figure 1 shows the Hornad catchment, topography of Margecany subcatchment, and location of gauging stations and the reservoir. The Margecany basin is a mountainous catchment, with elevations ranging from 339 to 1,556 m. The mean elevation of the basin is 670 m and the mean slope is about 18%. The stream order at the outlet is 6, and the maximum flow length is 90 km.

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 (Figure 1). Land cover data were obtained from 30 m Landsat-7 Enhanced Thematic Mapper satellite data, acquired on August 20th, 2000. The final land use map (Figure 2) for this study has a 100 $\times$ 100 m cell size and is composed of 6 different types of land cover: 49.8% of the basin is covered by forest (26.8% coniferous forest and 23.0% mixed-deciduous forest), 25.5% grassland and pasture, 22.8% agriculture areas, 1.8% urban area and about 0.1% water surfaces which are
mainly reservoirs. There are 10 different soil textures in the catchment. The dominant soil texture is loam, which covers about 42% of the basin, and sandy loam and silt loam about 24% and 23% respectively.

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 9 stations within the basin is 678 mm, ranging from about 640 mm in the valley to more than 1,000 mm in the mountains. The mean temperature of the catchment based on a 40-year period isothermal map is about 6°C. The annual potential evapotranspiration based on 10-years data of 1 station (Spisske Vlachy) is about 518 mm. 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 9 stations, temperature for 4 stations, PET at 1 station, and daily discharge data at 8 gauging stations. All these data are available for a 10-year period from 1991 to 2000. Daily discharge data at 8 locations are available inside the catchment, but only the Margecany station is used for model calibration.

Figure 1 Hydrologic network of Hornad catchment, topography of Margecany subcatchment, and location of gauging stations and Ruzin reservoir

Figure 2 Land use map of the Margecany watershed
Model simulation

Once the required data are collected and processed for use in the WetSpa model, 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 a stream when the upstream drained area is greater than 0.1 km². The threshold value for determining subcatchments is set to 1,000, by which 69 subcatchments are identified with an average subcatchment area of 16.6 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. The grid of hydraulic radius is generated with an exceeding frequency of 0.5 (2-year return period) resulting in an average hydraulic radius of 0.005 m for the upland cells and 1.5 m at the outlet of the main river 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, grids of root depth, interception storage capacity, and Manning’s roughness coefficient are reclassified from the land use grid, in which the Manning’s coefficient for channels is linearly interpolated based on the stream order grid with 0.055 m$^{2.3}$ s for the lowest order and 0.025 m$^{2.3}$ 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%. The results are shown in Figure 3a. From this figure it follows that non-forested and steeper areas have a very high potential runoff coefficient, whereas the forested and gentle slopes generate less surface runoff. The calculated average potential runoff coefficient is 0.43 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, mean travel time to the basin outlet as well as the standard deviation are generated, which enable us to calculate the IUH from each grid cell to the basin outlet. Figure 3b shows the estimated average flow time from each grid cell to the basin outlet; the flow time for the most remote area is around 62 h, while the mean travel time for the entire catchment is 25 h.

Results and discussion

Daily stream flow simulation

The 10-year (1991–2000) measured daily discharge data are used for model calibration and validation. The calibration process is performed manually for the global model parameters only, whereas the spatial model parameters remain as present. Initial global
model parameters are specifically chosen according to the basin characteristics as discussed in the documentation and user manual of the model (Liu and De Smedt, 2004). The simulation results are then compared to the observed hydrograph at Margecany 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 groundwater flow recession coefficient is adjusted by fitting the baseflow, which is separated from the observed hydrograph. The interflow scaling factor, which is sensitive for high flows, is adjusted for the recession part after flood peak. The 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. Initial soil moisture and active groundwater storage are adjusted by comparison of the hydrograph and water balance for the initial phase. The maximum active groundwater storage controls the amount of transpiration from the groundwater, and therefore can be adjusted by comparison of the flow volume during dry periods. Figure 4 gives a graphical comparison between observed and calculated daily flow at Margecany for the years 1991 and 1997. It shows 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 2.0 mm/d°C. The calibrated groundwater flow recession coefficient at Margecany is 0.0085 d⁻¹, and gives a good estimation for all base flows. 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, as the flow volume is 0.4% over-estimated, the Nash–Sutcliffe efficiency is 0.80, and the modified Nash–Sutcliffe efficiency is 0.81 and 0.82 respectively for high and low flows. This indicates that the model is able to simulate hydrologic 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. The model outputs also show that 7.8% of the precipitation is intercepted by the plant canopy, 84.5% infiltrates to the soil, 67.9% evaporates to the atmosphere, 23.5% recharges to the groundwater reservoir, and 31.3% becomes runoff, of which direct flow, interflow, and groundwater flow contribute 5.7%, 3.8% and 21.8% respectively. These values are reasonable in view of the catchment hydrological characteristics.

Figure 4 (a) Graphical comparison between observed and calculated daily flow at Margecany for the year 1991; and (b) for the year 1997
Hourly precipitation data in the summer of 2001 were obtained from SHMU for 5 stations within the catchment, together with hourly discharge data at Margecany station. Hourly evaporation data for this period were calculated based on monthly data measured at Spisske Vlachy (Figure 1) in the summer of 2001 and an empirical curve which was developed based on 10 years of daily observed PET data, presented in the documentation of the model (Liu and De Smedt, 2004). Figure 5a gives a graphical comparison between observed and calculated hourly flows at Margecany during the summer of 2001. Despite incompleteness of precipitation data and insufficient evaporation data, the simulated hydrograph shows a good agreement with the measured hydrograph.

Land use has a great influence on the rainfall runoff process. 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 land use changes on the hydrologic behavior of a river basin. For assessing the impact of land use, in particular forestation, the calibrated WetSpa model was applied for a reforestation scenario using the hourly data of the summer 2001. For the reforestation scenario it is assumed that the forest will increase by 50%. This is achieved by converting grassland to forest, because grassland is almost everywhere situated between forest and agriculture areas. Currently about half (49.8%) of the catchment is covered by forest, hence a 50% increase in forests will yield a forested area of 75.3% in total.

Based on this land use change scenario, the model was run to estimate the modified flows. It is worth mentioning how land use changes affect the simulation parameters in the model. After changing the land use map the following maps have to be recalculated: root depth, interception, Manning coefficient, depression storage, runoff coefficient, velocity, flow travel time, and the standard deviation of travel time. The changes in the flow travel time and the standard deviation consequently also change the IUH of each cell and the watershed IUH. Figure 5b gives the simulated flood hydrograph for the present and the reforestation scenario for a flood event that occurred in July 2001. The simulated peak discharge for the present land use is 84.8 m$^3$/s, and for the reforestation scenario this becomes 74.9 m$^3$/s, which means that the peak discharge is reduced by 12%. Investigation of peak discharges for the whole simulation period, shows that the reforestation scenario results in a reduction of 18% on average, while for small discharges the reduction is even about 34%. In addition to the difference in magnitude of the simulated peak discharges, differences in time to peak are also observed. The time to peak of the reforestation scenario is 20 hours longer than for the present land use. Additionally, the effects of land use change on flood volume, runoff composition, evapotranspiration and soil moisture can also be evaluated quantitatively from the model results. It is found that reforestation results in decreasing surface runoff and flood volume, but increasing

![Figure 5](https://iwaponline.com/wst/article-pdf/53/10/37/432047/37.pdf)
amounts of interflow and baseflow, as well as soil moisture and evapotranspiration. The magnitude of changes, however, differs from one storm to the other depending upon the antecedent soil moisture conditions.

Similar studies, using different models, by Lahmer et al. (2001) and Niehoff et al. (2002) also mentioned the importance of land use change and forestation on flood generation. Lahmer et al. (2001), for example, presented a forestation scenario assuming the conversion of all the arable land (about 66% of the total basin area) to forest that has reduced peak flow by 42.3%.

Conclusions
In this paper an attempt is made to outline a method for estimating flood runoff by using detailed basin characteristics together with meteorological data as an input to the WetSpa spatially distributed model. 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, soil type and their combinations are collected from the literature, which can be prepared easily using standard GIS techniques. The model is tested on the Margecany mountainous catchment in Slovakia with 10 years of observed daily rainfall and evaporation data. Good agreement with the measured hydrograph is achieved. 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. For assessing the impact of land use changes, the calibrated WetSpa model is applied for a reforestation scenario using the hourly data of the summer of 2001. A 50% increase of forests in the catchment shows only a moderate reduction in discharge. The results show the reforestation scenario reduces the discharge by 18% on average. The time to peak of the simulated hydrograph in the reforestation scenario becomes 20 hours longer than for the present land use.

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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References


