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HYDROLOGICAL MODELING WITH SPECIAL REFERENCE TO SNOW COVER PROCESSES

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Abstract. *A physically-based distributed hydrologic model was applied in this research. The river basin, or watershed, was discretized with a square grid, where each square carried morphological data about a portion of the watershed, the vegetation, the soil composition, the hydrogeological layer, and the like. The effect of weather stations was defined by Thiessen polygons, including correction for altitude. The hydrological model was continuous, with a one-day time step. It was partitioned into three reservoirs: vegetation, snow and soil. The snow reservoir was defined using the degree-day and temperature index methods. The hydrologic model was applied to the basin of the Banjska River, which is a tributary of the Južna Morava.*

Key words: *Hydrological modeling, snow accumulation, snowmelt.*

1. INTRODUCTION

The transformation of precipitation into runoff is a highly complex natural process such that modeling is rather intricate. A large number of input data are required for the hydrologic model, whose quality determines the simulated runoff. In physical models hydrological processes are represented by mathematical equations that describe the behavior of nature. However, often this behavior cannot be explained by physical laws and models have to resort to empirical dependencies of certain quantities.

This research analyzes a physically-based distributed model that requires the watershed to be discretized with a square grid where each square carries information about the morphological characteristics of a portion of the watershed, the vegetation, the soil composition, the hydrogeological layer, and the like. It is necessary to define the effect of weather stations in the watershed, generally determined by means of Thiessen polygons, including correction for elevation.

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Evaporation depends on numerous meteorological factors, such as air temperature, air humidity, vapor pressure, atmospheric pressure and radiation. The methods used to analyze the evaporation process are described in the paper. The evaporation process is treated in the model by the Hargreaves equation where the air temperature and radiation at the top of the atmosphere are used to determine potential evapotranspiration. The vertical soil profile is divided into two parts: the surface layer (topsoil) and the subsurface layer. The subsurface layer is comprised of four sub-layers. Actual evapotranspiration is comprised of the evaporation process (evaporation from plants and soil) and transpiration (water consumption by plants for their own needs). The model assumes that evaporation takes place only in the surface layer, while transpiration occurs in subsurface layers of the soil.

Surface runoff takes place in the surface layer. The coefficient of runoff in a watershed depends on the soil moisture, type of vegetation and type of soil. The SCS method is used to describe this process. Some of the precipitation is infiltrated into the soil and moves vertically down to an impermeable bed. The water that flows into deeper layers is used by plants for their own needs. The remainder of the water below the ground surface recharges groundwater.

The complex process of snowmelt is modeled in two ways. The first is based on the Temperature Index Method and the second on the Degree-Day Method, where snowmelt is a linear function of air temperature and solar radiation.

Water discharge at the point of outflow from the watershed is delayed due to water retention within the watershed and river network. The retention effect increases with increasing watershed size. The principle of linear reservoirs is applied to mathematically represent the water retention process in the model.

2. Hydrologic Cycle in Runoff Models

Parametric hydrology deals with the genesis of hydrological processes. The term "parametric" refers to the fact that hydrologic process parameters are studied. Parametric hydrology has two basic tasks:

- water budgeting in different areas and different time units and
- transformation of precipitation into runoff.

Diverse parametric hydrology methods are used. The basic water balance equation for a watershed or reservoir is (Bonacci, 2011):

$$U - I = \Delta V, \tag{1}$$

where U is the input of all waters into the considered area, I is the output of all waters from the considered area, and ΔV is the change in water volume in a given area and time interval.

A hydrologic year encompasses the period of a full hydrologic cycle. It varies and depends on the climate and geographic location. It generally lasts between 11 and 13 months. To facilitate hydrologic estimation, it is assumed that the hydrologic cycle lasts for 12 months. In the considered region, the hydrologic year begins on 1 October and ends on 31 August.

Components of the hydrologic cycle

Precipitation is transformed both underground and on the ground surface, but in shallow strata. The total runoff from a watershed is divided into surface runoff and base runoff. Base runoff is the sum of runoff along the ground surface and subsurface runoff from the thin initial layer of the soil. A small portion of precipitation that contributes to surface runoff is that which falls directly on the water table of a river. Base runoff is formed from a portion of the delayed subsurface flow and predominantly from water in the underground that flows along the direction of the maximum gradient of the groundwater level.

Water losses occur as a result of evaporation of water from the soil and vegetation, as well as water uptake by plants. When rain falls, a portion of the water is trapped by the leaves of plants and the remainder reaches the ground. For rain to reach the ground, the amount of precipitation needs to be greater than the capacity of the plants to trap the water on their leaves. This phenomenon is called interception and it varies during the hydrologic cycle; it is the greatest in the spring and summer when vegetation is plentiful. Interception is a part of evaporation from the soil and vegetation.

Some of the water that evaporates from the soil comes from snow. The difference between this physical process and evaporation from the vegetation cover is that there is a direct change from the solid to the gaseous (water vapor) state. This process is called sublimation.

Evaporation from the ground includes evaporation from the soil and from vegetation, such that interception and sublimation are parts of this process. Additionally, some of the moisture reaches the atmosphere during the course of transpiration, where plants take up water from the soil, use it and return it to the atmosphere. The summary process of evaporation from the ground surface and plant transpiration is called evapotranspiration. Heat is needed for evaporation from the ground surface, to convert water into vapor. The energy needed for molecules to transport to a gaseous state is called latent heat of evaporation (Plavšić, 2001). Evapotranspiration increases air humidity. When the air is saturated with water vapor, evaporation ceases. The point of saturation depends on the air temperature and atmospheric pressure. For evaporation to take place, there must be an air humidity deficit between the surface which evaporates and the air above it. The evapotranspiration rate that would be achieved if the ground surface along with vegetation had an unlimited source of moisture is called potential evapotranspiration. Fig. 1 is a schematic representation of the transformation of precipitation into runoff in all parts of the hydrologic process.

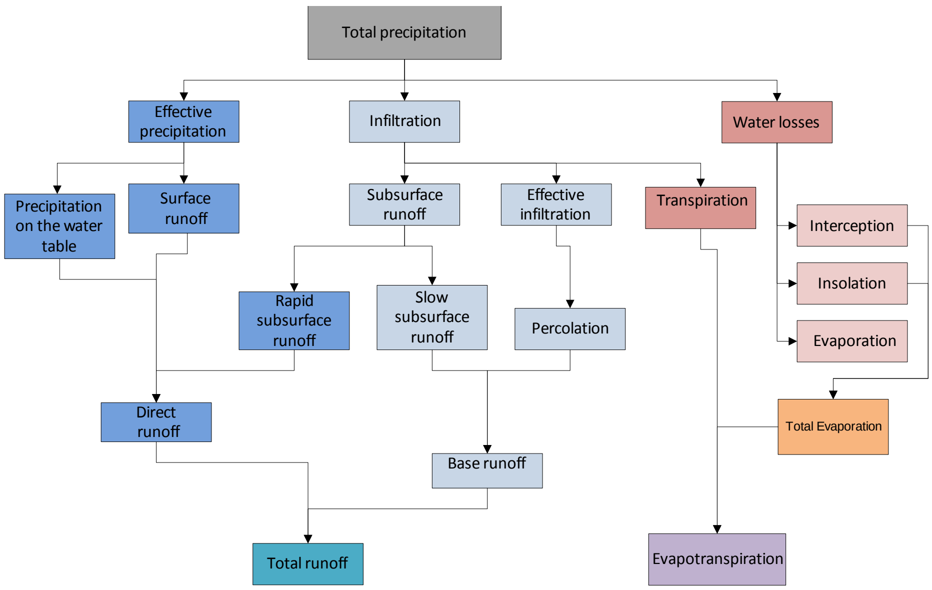

Fig. 1 Transformation of precipitation into runoff

The entire hydrologic cycle (see Fig. 1) is a closed cycle since the water that circulates through the system stays within it. This system functions owing to the fact that there is excess solar radiation that reaches the earth, relative to the radiation that leaves (Jovanović et al., 1990). The hydrologic cycle can be divided into several subsystems, such as the atmospheric subsystem, the ground surface subsystem, the subsurface system, the groundwater subsystem, the river subsystem and the ocean subsystem. Hydrology does not deal with the entire system; meteorology addresses processes that take place in the atmosphere and oceans.

In order to analyze the hydrologic cycle, its functioning needs to be described using a formal language. The formalization represents a set of mathematical relations that can be algebraic equations, linear and non-linear differential equations.

For continuous unsteady dynamic systems of the n-th order with a single input - $x(t)$ and a single output - $y(t)$, the mathematical model is a simple differential equation of the n-th order (Jovanović et al., 1990):

$$a_n(t)\frac{d^n y(t)}{dt^n} + a_{n-1}(t)\frac{d^{n-1} y(t)}{dt^{n-1}} + ... + a_0(t)y(t) = x(t). \tag{2}$$

If in the hydrologic model the model parameters $a_i(t)$ do not depend on $x(t)$ and $y(t)$ and their derivatives, the system is linear. Otherwise, the system is non-linear.

The hydrologic cycle is steady if the coefficients ai are constants, or if a_i=$a_i(t)$. If the coefficients ai are time-dependent, the system is unsteady.

With regard to distinctness, there are deterministic and stochastic models (Bonacci, 2011). To represent reality as accurately as possible, the model needs to simulate the random nature of variables. The stochastic model is a model that recognizes the random nature of inputs. A model that does not include a random component is deterministic in nature. The output of deterministic models is determined as soon as the inputs and their relationships are defined. Conversely, in a stochastic model the output is random, as are the inputs which constitute random variables. The deterministic model is only a special, simplified case of the stochastic model. The choice of model to be used will depend on whether the output is a single scenario derived from deterministic models or a set of possible scenarios resulting from stochastic models.

Based on the explanation of natural processes, hydrologic models are classified into (Bonacci, 2011): systemic (black box), empirical (gray box) and physical (white box).

Representatives of black box models are the rational method, unit hydrograph and transfer functions. The transformation of precipitation into runoff began in the 19th century using the rational method, where discharges are directly correlated with precipitation in the watershed, the size of the watershed and the runoff coefficient (Anderson et al., 2005). In this case the runoff coefficient may be considered as a model parameter that varies to reflect the local conditions in the watershed. The corresponding hydrograph has the shape of an equilateral triangle and is generally used for small, semi-permeable watersheds.

A considerable breakthrough in hydrological modeling was made in the 20th century. The best known discovery was the unit hydrograph, in an attempt to determine the temporal position of the hydrograph peak (Anderson et al., 2005). The unit hydrograph represents unit effective rainfall equally distributed across the watershed, obtained by subtracting from total precipitation the portion of precipitation infiltrated into the ground. The hydrograph is not triangular like that produced by the rational method; it corresponds to the perceived hydrograph normalized for unit precipitation (mm, cm). In this model the problem of determining the runoff coefficient remains; namely, the portion of total rainfall that represents effective rainfall needs to be established. This depends on prior soil moisture, watershed morphology, hydrogeological and pedological compositions of the soil, vegetation cover and other factors.

Recently, transfer functions (*TFs*) have been used to transform precipitation into runoff (Anderson et al., 2005). Transfer functions are common in many branches of science. In hydrology, continuous-time *TF* models are used. According to the explanation of natural processes, transfer functions represent black box models. The transfer function $h(t)$ is determined on the basis of known inputs $x(t)$ and outputs $y(t)$. It defines the transformation between the input rainfall hyetograph and the output hydrograph. The Fourier series is used to identify the hydrograph, where the inputs and outputs are represented as the sums of cosinusoidal and sinusoidal waves. The transfer function is defined using the convolution integral (Jovanović et al., 1990). In addition to this approach, transfer functions are derived using *Fourier* transforms in the frequency domain, polynomial approximation and *Z transform* (Jovanović et al., 1990).

When modeling the transformation of precipitation into runoff, the greatest challenge is the definition of effective precipitation in the watershed. Under same soil moisture conditions, double the rainfall produces double the runoff. However, if the same rain falls on dry soil, the runoff will be considerably smaller. This non-linear problem has partly been addressed by introducing simulation models. Models are divided into continuous and short-episode models (Jovanović et al., 1990). Models that simulate short episodes include those that address flood waves, discharge rates during dry periods and complex direct and base runoff. The models are comprised of parameters that are assessed or calibrated for a given watershed. In continuous models, the state of the soil reservoir is computed for each time step. Some of these models, like the Stanford Watershed Model, feature 30 parameters.

Based on their spatial anatomy, hydrologic models feature distributed and lumped parameters. Lumped hydrologic models include, for example, *SSARR*, *Stanford*, *Dawdy-ODonnell*, *Tank* and *HBV*. Representatives of distributed models are *ILR* and *SWAT*.

Considerable progress in hydrological modeling was achieved when distributed models were developed. Lumped models view the watershed as an integrated whole, attempting to correlate precipitation input and runoff output, without considering the spatial characteristics of the watershed. In contrast, distributed hydrologic models address watershed characteristics, such as morphology and climate, when they transform precipitation into runoff.

Hydrologic models can also reflect partially distributed data. The watershed, or river basin, needs to be divided into sub-basins, and then parameters specified for the sub-basins, which are a function of spatial characteristics. Later the outflows from the sub-basin propagate downstream, such that total discharges exhibit a dependency on sub-basin characteristics. This approach allows for sub-basin features to be taken into account, but each sub-basin is viewed as a homogeneous entity, which is not the case in nature. A sub-basin at a higher altitude will receive more precipitation than a low-altitude sub-basin. Snowmelt is also a function of altitude and is affected by the terrain and other factors that cannot be addressed by models with partially distributed parameters. The set of parameters for a model needs to be assessed or calibrated for each sub-basin individually.

Fully distributed models take all spatial data into account through spatial discretization of the sub-basin. Spatial discretization elements may be of a regular structure comprised of a square grid and the grid that divides the watershed can also be an irregular triangular structure. Each element of the sub-basin has its own parameter values. Many models include multiple vertical layers that allow for 3D visualization of water movement.

The *ILR* model applies a genetic approach to translate rainfall to the point of outflow from the watershed (Jovanović et al., 1990). The watershed is divided by isochrones based on which a correlation is established between the concentration time and the surface area of the watershed. The linear reservoir is used in the model to simulate the retention capacity of the watershed, manifested in hydrograph maintenance at the point of outflow. The watershed is divided by means of a square grid, such that each square carries data that define the non-uniformity of precipitation and heterogeneous characteristics of the watershed from the viewpoint of water losses during a downpour. The model features three parameters: average runoff coefficients for the considered rain episode, river basin concentration time (isochrone map) and linear reservoir constant.

The concept of linear reservoirs is used for horizontal transformation of runoff in hydrological modeling. According to the rational theory, the hydrograph is modeled by means of a triangle. Apart from simplifying the actual appearance of the hydrograph, such modeling leads to maximum discharges occurring prior to real discharges and exhibiting higher values. This problem is a result of disregarding the effect of water retention in the watershed and river network (Jovanović et al., 1990).

The *SWAT* model was developed on the basis of digital terrain models and hydroinformatic methods. Digital terrain models are usually defined by a grid of squares with the same altitude (Stojković, 2012). This grid, applying the graph theory, provides information about flow directions in each part of the watershed and about square grid accumulation. The river network is formed based on the initial condition of the number of accumulated squares, and then the river course is segmented. The watershed is delineated at the most downstream parts of the segments. Spatial modeling of the terrain is based on a grid of hydrological response units (*HRUs*), whose size depends on the required accuracy of the model. Each *HRU* carries information about the altitude, soil composition, vegetation cover and other characteristics. The effects of water retention in the watershed (surface reservoirs) and the river network are addressed by means of linear reservoirs.

3. Material and Methods

The hydrologic model was applied to the Banjska River Basin. To simulate discharge, a distributed hydrologic model was used, based on a rectangular grid of *HRUs*, 250x250 *m*, for which the basic components of the vertical water balance were computed. Using a digital terrain model, the vegetation cover and the pedological and hydrogeological compositions of the soil, an .mdb database with the attributes of the given layer was created for each *HRU*.

The horizontal water balance was assessed at the locations of hydrological stations and hydraulic structures. Fig. 2a shows the sites of the Vranjska Banja Hydrological Station (HS) and the Prvonek Reservoir. An illustration of the *HRU* grid is provided in Fig. 2b.

Fig. 2c shows the hydrogeological information used to adopt parameters such as flow velocity and porosity of subsurface layers. The pedological composition of the soil determined the saturation flow velocity and the porosity of the surface layer of the ground (see Fig. 2d). The parameters that affect evaporation and surface runoff were defined based on the vegetation cover (see Fig. 2e).

The model parameters were homogenized commensurate with their extent in each of the *HRUs*, to generate uniquely-defined parameters of the *HRU* for all soil and vegetation characteristics.

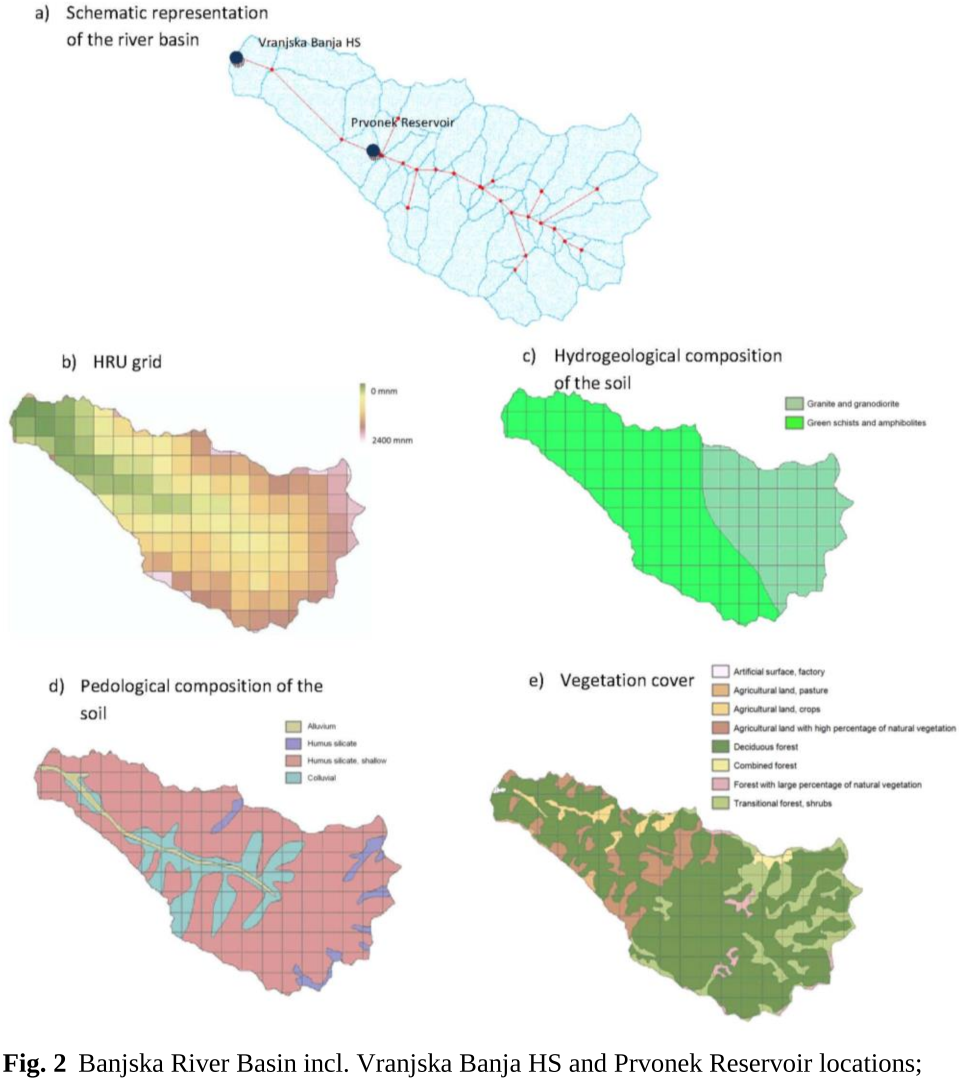

Fig. 2 Banjska River Basin incl. Vranjska Banja HS and Prvonek Reservoir locations; division of the river basin into HRUs (b) with the hydrogeological composition of the soil (c), pedological composition of the soil (d) and vegetation cover (e).

3.1. Meteorological inputs

Meteorological inputs need to be defined in order to generate a hydrologic model with a daily time step. Reference precipitation and temperature levels are computed for each *HRU*. Precipitation increases with increasing altitude, while temperature decreases. The calculation of reference precipitation levels in the *HRUs* is related to rain gauge stations grouped by means of Thiessen polygons. Precipitation and temperature gradients need to be defined based on data available from weather stations in the river basin, or, in other words, a correlation needs to be established between the increase (decrease) in meteorological quantities with altitude variation. Reference precipitation and temperature levels for each *HRU* are computed using the following formulas (Neitsch et al., 2011; Prodanović et al, 2009):

$$R_{HRU}(i) = \sum_{k=1}^{n} \left\{ R_{gauge}(i) + (EL_{HRU}(i) - EL_{gauge}) \cdot \frac{p_{laps}}{1000} \right\}, \tag{3}$$

$$T_{HRU}(i) = \sum_{k=1}^{n} \left(T_{gauge} + \left(EL_{HRU}(i) - EL_{gauge} \right) \cdot \frac{t_{laps}}{100} \right), \tag{4}$$

where: R_{HRU}, T_{HRU} are the reference values of precipitation and temperature for the *HRU* (*mm*, ^{o}C); R_{gauge}, T_{gauge} are precipitation and temperature levels recorded by the weather station (*mm*,^{o}C); EL_{HRU} is the average altitude of the considered *HRU* (*m.a.s.l.*); EL_{gauge} is the altitude of the weather station (*m.a.s.l.*; and p_{laps}, t_{laps} are the precipitation and temperature gradients depending on altitude (*mm/km*, $^{o}C/km$).

A time counter i=1,2,…n (number of simulated days) was applied in Equations (3) and (4). For the Banjska River Basin, one weather station (Vranje, alt. 432 m) and one rain *gauge* station (Kriva Feja, alt. 1100 m) were used. Precipitation and temperature gradients were computed for the extended area, such that p_{laps}=1.05 *mm/km* i t_{laps}=-3.45 $^{o}C/km$.

3.2. Vertical water movement

The soil moisture level is based on the initial equation of the hydrologic model (Neitsch et al., 2011):

$$SW_t = SW_0 + \sum_{i=1}^{t} (R(i) - Q_{surf}(i) - W_{perc}(i) - E_s(i) - E_{sub}(i) - W_{up}(i) - E_{can}(i)), \tag{5}$$

where SW_t is the ultimate soil moisture, SW_0 is the initial soil moisture, R is the daily precipitation total, Q_{surf} is the daily direct runoff, W_{perc} is the daily base runoff, E_s is the daily evaporation from the ground, E_{sub} is the daily evaporation from the snow (sublimation), W_{up} is the daily water uptake by plants (transpiration), and E_{can} is the daily interception rate. It is important to note that *mm* water units were used for all hydrologic cycle parameters.

3.2.1. Potential evapotranspiration

Potential evapotranspiration is calculated by means of the empirical *Hargreaves* method (Samani, 2005). This method requires minimal input data (i.e. minimum and maximum daily air temperatures and top-of-atmosphere radiation). It is defined by the following equation:

$$\lambda E_0(i) = 0.0023 H_0 (T_{\max}(i) - T_{\min}(i))^{0.5} (T_{av}(i) + 17.8), \tag{6}$$

where E_0 is the daily potential evapotranspiration (*mm*), λ is the latent evaporation heat ($MJkg^{-1}$), H_0 is the top-of-atmosphere radiation (MJm^{-2}), and T_{max},T_{min},T_{av} are the maximum, minimum and average air temperatures (^0C), respectively.

The latent evaporation heat determines the energy flow when water is transformed from a liquid to a gaseous state. It is defined as (Plavšić, 2012):

$$\lambda(i) = 2.501 - 2.370 T_{av}(i)/1000. \tag{7}$$

Total precipitation R (*mm*) is then divided into rainfall R' (*mm*) and snowfall, applying the criterion that snow is formed at *HRU* air temperatures below 0^0C.

3.2.2. Interception

Rainfall retention on plants affects infiltration into the soil, surface runoff and evaporation from the ground. The water trapped on plants reduces the erosive energy of the rain that would fall if there were no vegetation cover. The effect of reducing erosion by interception increases with increasing density of the vegetation cover, with plant species that trap relatively large amounts of rain. The interception rate is the highest in river basins featuring dense forests, where it can have a significant impact on the water balance. Interception has been measured in North Carolina (USA), in two types of dense forests, and the results were as follows (Jovanović et al., 1990):

Table 1 Interception measured in North Carolina (USA)

	Precipitation (mm)			
	10	20	50	70
Type of forest	Interception (mm)			
Coniferous	2.5	4.3	4.8	5.6
Deciduous	1.5	2.0	3.1	3.4

Table 1 leads to the conclusion that interception is a non-linear process that is a function of the extent of rainfall. As the amount of rainfall increases, the increment of water trapped on the leaves decreases to the maximum water retention capacity of the leaves.

The daily interception rate is the amount of water that plants can trap on their leaves and which later evaporates. This water does not take part in the formation of surface runoff or in infiltration. The maximum daily interception rate is (Neitsch et al., 2011):

$$can_{day}(i) = can_{\max}(i) \frac{LAI(i)}{LAI_{\max}}, \tag{8}$$

where can_{day} (*mm*) is the maximum daily amount of water trapped by plants, *LAI* (m^2m^{-2}) is the leaf surface index that varies during the growing season and depends on the type of vegetation, and LAI_{max} (m^2m^{-2}) is the maximum value of this index.

In the first water budgeting step, the total precipitation in a given reservoir is reduced by the amount trapped by vegetation (Neitsch et al., 2011):

$$R_{\text{int}}(i) = R_{\text{int}}(i-1) + R'(i) - can_{day}(i), \tag{9}$$

where $R_{int\ (i-1)}$ (mm) is the previous state of the reservoir and R' (mm) is the instantaneous rainfall reduced by the instantaneous amount trapped by plants can_{day} (mm).

The condition is that total evaporation E_{can} (mm) from this reservoir cannot be greater than the potential evapotranspiration E_0 (mm):

$$E_{can}(i) \leq E_0(i), \tag{10}$$

where $R_{int\ (i-1)}$ (mm) is the previous state of the reservoir and R' (mm) is the instantaneous rainfall reduced by the instantaneous amount trapped by plants can_{day} (mm).

Potential evapotranspiration E_0 is then reduced by the interception rate E_{can}:

$$E_{01}(i) = E_0(i) - E_{can}(i), \tag{1}$$

To compute the relationship between potential evaporation from the ground and vegetation E_s (mm) and potential water uptake by plants (transpiration) E_t (mm), it is necessary to calculate cov_{soil} for each day of the year according to the following equation (Vasilić et al., 2012):

$$\text{cov}_{soil}(i) = \exp(-0.4 \cdot LAI(i)). \tag{2}$$

3.2.3 Snow cover

Information about water reserves in the river basin is needed to simulate runoff and predict inflow under snow conditions. Three methods are used to model snow accumulation and snowmelt (Kang, 2005):

- *The Degree-Day Method*,
- *The Temperature Index Method*, and
- *The Energy Balance Method*.

The Degree-Day Method is based on the correlation between snowmelt and air temperature. This correlation can be expanded to include the effect of solar radiation. The time scale can be less than one day, to address air temperature fluctuations during the day. In such cases the method is called the Degree-Hour Method.

The Temperature Index Method is similar to the Degree-Day Method and is adapted to the river basin by selecting a snowmelt temperature that may differ from 0 ^0C. HBV, SRM, UBC, HYMETl and SWAT models reflect such an approach.

The Energy Balance Method focuses on energy exchange between air and snow temperatures. In the simplified approach, convection (appreciable heat), condensation, radiation and rainfall take part in the snowmelt energy balance (Jovanović et al., 1990):

$$Q = Q_{\text{convection}} + Q_{\text{condensation}} + Q_{radiation} + Q_{rain}. \tag{3}$$

Snowmelt heat is transferred from the atmosphere to the snow cover through convection $Q_{convection}$ ($mm\ hour^{-1}$). Convection depends on air temperature, wind speed and latent snowmelt heat. Given that atmospheric turbulence contributes to the transport of humid air to the surface of the snow cover, the heat that traces to humid air condensation takes part in snowmelt proportional to the wind speed and water vapor gradient – $Q_{condenzation}$ (*mm hour*$^{-1}$). An important participant in snowmelt is the absorbed net radiation that can be determined based on meteorological monitoring data (Jovanović et al., 1990). The rate of snowmelt by radiation $Q_{radiation}$ (*mm hour*$^{-1}$) is defined by:

$$Q_{rad} = R_n / L, \tag{4}$$

where L is the latent snowmelt heat, or the energy delivered by the snow when it transforms to a liquid state. Raindrops contribute to snowmelt Q_{rain} (*mm hour*$^{-1}$) by transferring heat energy to the snow. Snowmelt by raindrops is affected by the amount of rainfall and air temperature.

When the snowmelt energy balance method was applied at different altitudes in the Alps, it showed that net radiation had the greatest impact on snowmelt. Convection was also a significant factor, while the impact of raindrops was minor (Jovanović et al., 1990).

The physical process of snowmelt relates to one portion (one point) of the river basin, but if the entire river basin is considered, the snowmelt process is rather complicated. One of the factors that affects snowmelt is the ability of the snow cover to retain and release water (Jovanović et al., 1990). Since runoff begins when the water retention capacity of the snow is reached, the beginning of melting is delayed.

The snowmelt process was examined in the hydrologic model in two ways. The first method was developed by the NWSRFS (National Weather Service River Forecast System) in the US and constitutes one of the Temperature Index Methods. According to this method, snowmelt depends on air temperature, snow temperature, snowmelt coefficient, and snow cover in the river basin (Melloh, 1999).

Snow temperature is a function of the average temperature during the day T_{av} (^0C) and the snow temperature during the previous day $T_{sno(i-1)}$(^0C) (Melloh, 1999):

$$T_{sno} = T_{sno}(i-1)(1-\lambda_{sno}) + T_{av}(i)\lambda_{sno}, \tag{5}$$

where λ_{sno} is the weight coefficient (effect of previous day snow temperature) between 0 and 1. The present model used 0.5.

Snowmelt is calculated as a linear function of the difference between the average snow temperature T_{sno}(^0C) and maximum air temperature $T_{max.dn}$ (^0C) with the snowmelt temperature T_{melt}(^0C) (Melloh, 1999):

$$SNO_{melt}(i) = b_{mlt}(i) sno_{\text{cov}} \left[\frac{T_{sno}(i) + T_{\max.dn}(i)}{2} - T_{melt} \right], \tag{6}$$

In Equation (16), the snowmelt coefficient b_{mlt} (*mm* $^0C^{-1}$ *day*$^{-1}$) is calculated on the basis of characteristic values of the coefficients b_{mlt6} and b_{mlt12} and the ordinal number of the day of the year - N_{day}. The adopted snowmelt temperature is T_{melt}=0 ^{0}C. The coefficient b_{mlt} is derived from (Melloh, 1999):

$$b_{mlt} = \frac{b_{mlt6} + b_{mlt12}}{2} + \frac{b_{mlt6} - b_{mlt12}}{2} \sin\left(\frac{2\pi}{365}(N_{day} - 81)\right). \tag{7}$$

The coefficients b_{mlt6} and b_{mlt12} are related to the beginning of summer (21st June) and beginning of winter (21st December), respectively. The following values are adopted in the model: b_{mlt6}=2.2 $mm\ {}^0C^{-1}\ day^{-1}$ and b_{mlt12}=0.6 $mm\ {}^0C^{-1}\ day^{-1}$.

The second, alternative method applied to analyze snowmelt is the *Degree-Day Method*. Contrary to the previous approach, where snow accumulations begin at 0^0C (temperature of storing snow), this method provides for snow storing at a higher temperature (Maho et al., 2011). It is assumed that snow accumulations are formed at temperatures lower than 3.1^0C (Mutreja, 1986). Given that snow is created in the higher reaches of the atmosphere, where the temperatures are lower than those on the ground surface, snow is known to fall on the ground at positive temperatures. To address this phenomenon, the potential snow cover coefficient – a (see Fig. 3) is introduced. For average daily temperatures less than or equal to zero, this coefficient is 1 (i.e. all precipitation at these temperatures is snowfall). For temperatures less than 3.1 ^0C, some of the precipitation is snow and the remainder is rain (Mutreja, 1986). The distribution of rain and snow is assumed to be linear, according to the following equation:

$$a = 100 - 32.3T_{av}. \tag{8}$$

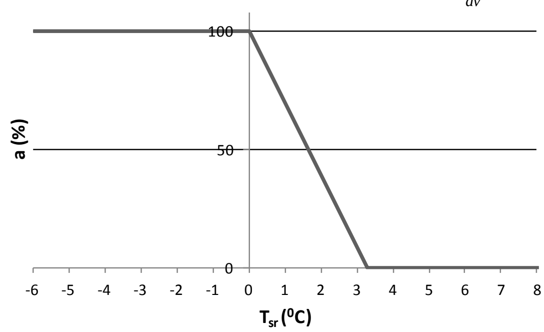

Fig. 3 Potential snow cover coefficient a (%)

The first step in snow accumulation is (Shaffe et al., 2011):

$$\begin{aligned} &SNO_i = R_{HRU} \cdot a/100 + SNO_{i-1},\ a = 100 - 32.3T_{sr.dn}\ (0 \le T_{av} \le 3.1\,), \\ &SNO_i = R_{HRU} + SNO_{i-1},\ (T_{av} < 0\,), \\ &SNO_i = SNO_{i-1},\ (T_{av} > 3.1\,), \end{aligned} \tag{9}$$

The standard Degree-Day Method is a linear function of snowmelt and temperature. Snowmelt is defined by the expanded Degree-Day Method that includes the effects of air temperature and solar radiation (Melloh, 1999):

$$SNO_{melt}(i) = T_{av}(i) \cdot c_T + H_0(i) \cdot c_H\,, \tag{20}$$

where T_{av} is the average daily temperature and H_0 is the net top-of-atmosphere radiation. The coefficients c_T ($mm\ day^{-1}\ {}^0C^{-1}$) and c_H ($mm\ day^{-1}\ (W\ m^{-2})^{-1}$) define the rate of snowmelt relative to unit variation in air temperature and solar radiation. The recommended value for the effect of solar radiation is c_H=0.26 $mm\ day^{-1}\ (W\ m^{-2})^{-1}$ (Shaffe et al., 2011), while the recommended values for the temperature effect are in the range 1-5 *mm day*$^{-1}$ $^{0}C^{-1}$. The value of the parameter used in the present case is c_T= 1.5 *mm day*$^{-1}$ C^{-1}.

The snow cover sno_{cov} needs to be calculated in order to compute the rate of snowmelt, based on the parameter SNO_{100} (*mm*) that corresponds to the snow depth at which 100% of the *HRU* is covered with snow (Vasilić et al., 2012):

$$sno_{\mathrm{cov}} = \frac{SNO}{SNO_{100}}. \tag{10}$$

It is assumed that $SNO_{100}=5$ *mm* results in full *HRU* coverage. Calculations are performed applying both snowmelt modeling methods.

Following snowmelt calculations, a correction is made for the state of the snow cover:

$$SNO_1(i) = SNO(i) - SNO_{mlt}(i). \tag{11}$$

Sublimation is calculated only if there is snow in the river basin after snowmelt, or, in other words, if SNO_1 (*mm*) is greater than zero. If there is snow, the parameter cov_{soil} from Equation (12) is also corrected, such that the adjusted value of cov_{soil_1} cannot be less than 0.5.

Actual sublimation amounts to (Neitsch et al., 2011):

$$E_{sub}(i) = E_{01}(i) \cdot \mathrm{cov}_{soil\_1}(i). \tag{12}$$

Finally, the remaining amount of snow for this reservoir is computed as:

$$SNO_2(i) = SNO_1(i) - E_{sub}(i). \tag{13}$$

3.2.4 Topsoil

Soil is divided into two parts. Surface runoff and evaporation from vegetation take part in the topsoil. The thickness of the topsoil is assumed to be 45 *mm*. Soil moisture can be related to the depth of the water table relative to the depth of the soil per unit of surface area. The field capacity (*FC*) is the amount of moisture in the soil when the capillary and gravity forces are in a state of equilibrium. Excess water on the ground surface is gravitational water that flows into deeper layers and recharges groundwater. The (permanent) wilting point (*WP*) represents the minimum soil moisture which the plants need in order not to wilt, below which transpiration ceases. The maximum storage capacity of the soil (*STO*), which depends on the porosity of the soil, is the moisture content of the topsoil where all pores are filled with water. Soil moisture cannot be greater than the maximum storage capacity of the soil.

The *SCS* runoff equation is applied in the present model. It is an empirical method in use since 1950 to study the correlation between rainfall and runoff in US mountainous watersheds. The model was developed to adapt to different types of soil and vegetation. The *SCS* equation is (Schiariti, 2012):

$$Q_{surf}(i) = \frac{(R(i) + SNO_{mlt}(i) - I_a(i))^2}{R(i) + SNO_{mlt}(i) - I_a(i) + S(i)}, \tag{14}$$

Q_{surf} (*mm*) is the surface runoff during the day, *R (mm)* is the rainfall during the day, SNO_{mlt} (*mm*) is the amount of snowmelt on that day, I_a (*mm*) is the initial abstraction (initial runoff losses), *STO (mm)* is the maximum soil moisture, and *S* (*mm*) is the potential (maximum) water deficit.

The potential soil moisture deficit is calculated based on *CN* as follows (Schiariti, 2012):

$$S = 25.4 \times \left(\frac{1000}{CN} - 10 \right). \tag{15}$$

The runoff curve number (*CN*) is a combination of the hydrological soil group and the type of use of the soil (Prohaska, 2003). This parameter is an indication of the runoff potential of the soil; the higher the *CN,* the greater the runoff potential. In addition to land use, to determine *CN* the hydrologic group of the soil needs to be known. There are four such groups (A, B, C, D).

It has been determined experimentally that initial abstraction is approximated by the equation I_a=0.2*S*; it represents the initial loss in the formation of surface runoff (Prohaska, 2003). Surface runoff occurs only when the sum of daily precipitation and snowmelt is greater than the initial abstraction or initial losses.

The adopted *CN* in Equation (26) is related to a terrain gradient of 5%, while a correction is needed based on the actual terrain slope *ST* (%) of the *HRU*. The corrected value of the parameter S_{2s} and consequently CN_{2s} are computed as (Neitsch S. et al., 2011):

$$S_{2s} = S \cdot \left[1.1 - \frac{ST}{ST + \exp(3.7 + 0.02117 ST)} \right], \tag{16}$$

$$CN_{2s} = 1000 \cdot \left[\frac{S_{2s}}{25.4} + 10 \right]^{-1}. \tag{17}$$

The *SCS* method identifies three conditions of previous soil moisture. CN_1 defines dry soil, at wilting point. CN_2 stands for average soil moisture, while CN_3 denotes wet soil whose soil moisture corresponds to field capacity. The *CN* for dry soil (29) and wet soil (30) is (Neitsch et al., 2011):

$$CN_1 = CN_{2s} - 20 \cdot \frac{100 - CN_{2s}}{100 - CN_{2s} + \exp(2.533 - 0.0636 \cdot (100 - CN_{2s}))}, \tag{18}$$

$$CN_3 = CN_{2s} \cdot \exp(0.00673 \cdot (100 - CN_{2s})). \tag{19}$$

Characteristic values of the coefficients CN_1 and CN_3 can be determined prior to simulation, but the actual *CN* is established based on the moisture of the soil reservoir by interpolation between the values of CN_1 and CN_3 (Stanić et al., 2011):

$$CN'(i) = CN_1 + \frac{SW(i-1) - WP}{FC - WP}(CN_3 - CN_1). \tag{20}$$

CN' cannot be less than CN_1 or greater than CN_3. In Equation (31), *SW*(*i-1*) is the state of the soil reservoir on the previous day expressed in *mm* of water.

Once calculated, *CN'* is used and Equation (26) applied to again compute the potential storage capacity of the soil *S* (*mm*), as well as the initial abstraction I_a (*mm*). According to the *SCS* method, initial abstraction (initial loss) includes interception, which is computed separately in the model. That is why initial abstraction is reduced by the computed interception E_{int} (*mm*) (Stanić et al., 2011):

$$I_a(i) = 0.2 \cdot S - E_{int}(i). \tag{21}$$

The next step is to estimate the state of the surface reservoir after runoff:

$$SW_1(i) = SW(i-1) + (R(i) + SNO_{mlt}(i) - Q_{surf}(i)). \tag{22}$$

If the new state of SW_1 (*mm*) is greater than the maximum capacity of the reservoir *STO* (*mm*), the amount of runoff needs to be increased by the difference between the state of the reservoir and the maximum capacity of the reservoir (Stanić et al., 2011):

$$Q'_{surf}(i) = Q_{surf}(i) + SW'(i) - STO. \tag{23}$$

Potential evaporation from the soil E_s (*mm*) and potential transpiration E_t (*mm*) are computed as follows:

$$\begin{aligned} E_s(i) &= E_{01}(i) \cdot \mathrm{cov}_{soil}, \\ E_t(i) &= E_{01}(i) - E_s(i). \end{aligned} \tag{24}$$

Evaporation from the soil E_s needs to be reduced by the amount of energy spent on sublimation E_{sub}:

$$E_{s1}(i) = E_s(i) - E_{sub}(i). \tag{25}$$

Actual evaporation E_{s2} is limited by the soil moisture, such that if the amount of moisture in the topsoil is greater than the amount of moisture at field capacity (*FC*), evaporation from the soil corresponds to potential evaporation (E_{s1}). If the soil moisture is less than field capacity, then evaporation is equal to (Stanić et al., 2011):

$$E_{s2}(i) = E_{s1}(i) \cdot \exp\left(\frac{2.5(SW_2(i) - FC)}{FC - WP}\right). \tag{26}$$

To prevent "drying out" of the soil reservoir, the following constraint is introduced (Stanić et al., 2011):

$$E_{s2}(i) \leq 0.8 \cdot (SW_2(i) - WP). \tag{27}$$

Consequently, evaporation cannot be greater than 80% of the difference between the state of the soil moisture (SW_2) and the soil moisture at wilting point (*WP*). After actual evaporation is calculated, the state of the surface reservoir (SW_3) needs to be reduced by E_{s2}.

The final component of the water balance that should be calculated relative to the topsoil is percolation into deeper layers W_{perc} (Neitsch et al., 2011):

$$W_{perc}(i) = (SW_3(i) - FC) \cdot \left(1 - \exp(\frac{-\Delta t}{TT_{perc}}\right), \tag{28}$$

where W_{perc} (*mm*) is the percolation into deeper layers, while SW_3 (*mm*) is the soil moisture of the surface reservoir after reduction for actual evaporation. Percolation will take place only if the soil moisture is greater than that at field capacity (*FC*). The time step is set at Δt of one day. The percolation time is defined as:

$$TT_{perc} = \frac{STO - FC}{K_{sat}} \tag{29}$$

Equation (40) shows that the percolation time depends on the flow velocity in the topsoil (saturated flow velocity), and that the duration is proportional to the water depth in the topsoil, above the soil moisture defined by field capacity. The new state of the surface reservoir (SW_4) is reduced by the amount of percolated water (W_{perc}).

3.2.5 Subsurface soil

The amount of water that will be infiltrated into the soil depends on a large number of factors, above all the condition of the soil on the ground surface, the vegetation, the type of soil and its characteristics such as porosity and hydraulic conductivity, as well as the current level of soil moisture (Plavšić, 2001).

The topsoil comprised of a mixture of solid particles, water and gases is called the aeration zone. At depths immediately above the water table there is a saturated layer of soil where all the voids between solid particles are filled with water. This layer is called the capillary fringe. The thickness of the capillary fringe varies depending on the composition of the particles, from several centimeters in sandy soil to several meters in clayey soil (Jovanović et al., 1990).

Groundwater flow from each portion of the river basin (*HRU*) is simplified in the model, such that the assumed groundwater flow direction corresponds to the direction of the shortest distance between the given *HRU* and the point of outflow. The horizontal flow velocity is defined by the hydraulic conductivity of the subsurface layer of soil that corresponds to the given *HRU*.

In addition to percolation, transpiration (water consumption by plants) takes place in subsurface layers. Another assumption is introduced, that water uptake is proportional to the thickness of the layer and that if the layers are of equal thickness the water uptake by layer will be equal.

The water that has percolated from the upper layer $W_{perc,l-1}$ *(l-1)* reaches the subsurface layer (*l*), such that in the subsurface layer:

$$SW_l(i) = SW_l(i-1) + W_{perc,l-1}(i). \tag{30}$$

where $SW_{l(i-1)}$ (*mm*) is the moisture of the subsurface layer on the previous day.

Then percolation from this layer and the new state of the reservoir are computed from (Stanić et al. 2011):

$$W_{perc,l}(i) = (SW_l(i) - FC_l) \cdot (1 - \exp(\frac{-\Delta t}{TT_{perc}}), \tag{31}$$

$$SW_{l2}(i) = SW_{l1}(i) - W_{perc,l}(i). \tag{32}$$

where TT_{perc} is the percolation time from the subsurface layer, calculated in the same way as in Equation (40), except that here the flow velocity in the subsurface layer is used.

The potential water uptake by plants in a layer is proportional to the thickness of the layer (Stanić et al., 2011):

$$W_{up,l}(i) = \frac{E_t(i)}{\sum_i D_{s,i}} D_{s,l}. \tag{33}$$

where $D_{s,l}$ (*cm*) is the layer thickness and E_t (*mm*) is the potential transpiration.

On the other hand, actual transpiration is limited by the soil moisture, such that actual transpiration amounts to (Stanić et al., 2011):

$$W_{up,l_1}(i) = W_{up,l}(i) \cdot \exp\left[5 \cdot \left(\frac{SW_{l2} - WP_l}{0.25(FC_l - WP_l)} - 1\right)\right], \tag{34}$$

Actual transpiration ($W_{up,2l}$) is further limited by the capacity of the reservoir and cannot be greater than the difference SW_{l2}-WP_l. The new state of the subsurface reservoir less transpiration is:

$$SW_{l3}(i) = SW_{l2}(i) - W_{up,l2}(i). \tag{35}$$

3.3 Concentration time

The concentration time is the time that elapses from the time of rainfall anywhere in the watershed to the time of its contribution to the outflow from the watershed. The concentration time is divided into the time during which the water moves across the watershed T_{ov} and the time when it moves along the river channel T_{hp}. This is described by the equation (Neitsch et al., 2011):

$$T_{con} = T_{ov} + T_{hp}. \tag{36}$$

The concentration time in the watershed is derived from:

$$T_{ov} = \frac{L_{slp}^{0.6} \cdot n^{0.6}}{18 \cdot slp^{0.3}}, \tag{37}$$

where L_{slp} is the travel distance across the watershed (*km*), *slp* is the terrain gradient of the watershed (mm^{-1}) and *n* is the *Manning* roughness coefficient of the watershed.

The concentration time from the beginning of the river network to the point of outflow is calculated from:

$$T_{hp} = \frac{0.62 \cdot L \cdot n_{hp}^{0.75}}{Area \cdot 0.125 \cdot slp_{hp}^{0.375}}, \tag{38}$$

where L (km) is the central length of the river course, slphp (mm^{-1}) is the average gradient of the river course, nhp is the Manning-roughness coefficient of the river course, and *Area* (km^2) is the surface area of the watershed.

The concentration time of groundwater flow from the *HRU* to the given observation points is calculated according to the simplified formula:

$$T_{C,GW} = \frac{L_{GW}}{k_{sat,ot}}, \tag{39}$$

where L_{GW} (m) is the shortest length of the flow from the part of the watershed to the point of outflow and $k_{sat,ot}$ is the coefficient of water flow through the hydrogeological layer (mm/day).

3.4 Horizontal movement of water

Horizontal movement of water occurs on the assumption that there is a linear correlation between the state of the water reservoir in the watershed and the discharge at the point of outflow (Vasilić et al., 2012). This assumption introduces the effect of water retention in the watershed and river network.

The linear reservoir equation is (Vasilić et al., 2012; Bonacci, 2011):

$$S = kQ_d, \tag{40}$$

where Q_d is the discharge at the point of outflow, S is the state of the reservoir (water stored in the watershed) and k is the linear reservoir parameter.

The rate of variation in the linear reservoir volume is the difference between the incoming flow (Q) and the flow from the reservoir (Q_d):

$$Q - Q_d = \frac{dS}{dt}. \tag{41}$$

Equation (52) is a continuity equation. Using Eqs. (51) and (52), or applying the linear correlation between the state of the reservoir (S) and the outflow (Q_d), the following equation is derived:

$$Q - Q_d = k\frac{dQ_d(t)}{dt}. \tag{42}$$

The linear reservoirs are inter-connected in a branching pattern which stands for a simplified/aggregated hydrographic network (Stanić et al., 2011). It is aggregated such that a link between an upstream and downstream observation point in the model is represented by a linear reservoir. In the present case, the concentration time between them (k) is calculated applying Eqs. (47) and (50).

Integration of the equation for the linear reservoir for the period Δt yields the outflow from the linear reservoir(Q_d) as a function of inflow (Q) and the previous state (S_0) (Stanić et al., 2011):

$$Q_d = S_0(1-e^{-\Delta t/k}) + Q\left(1-\frac{1-e^{-\Delta t/k}}{\Delta t/k}\right). \tag{43}$$

Equation (54) is set for each *HRU* and thus a system of linear equations is generated and solved for unknown discharges from the linear reservoirs at the observation points. The calculation is repeated for each time step *Δt*, which is one day in the present case.

4. Result and Discussion

The presented hydrologic model was applied to the Banjska River Basin. The Banjska River is located in southern Serbia and is a tributary of the Južna Morava. Its mouth is in the vicinity of the City of Vranje and its river basin holds a multi-purpose reservoir, Prvonek.

The weather station at Vranje was used for the purposes of the hydrologic model. The station was commissioned in 1965 and is situated at an altitude of 432 *m.a.s.l.* The Vranje weather station provided daily air temperatures (average, minimum and maximum). The rain gauge station at Kriva Feja was part of the Lake Vlasina monitoring system from 1951 to 1991. It is located at an altitude of 1100 *m.a.s.l.*, in the immediate vicinity of the Banjska River Basin. Daily precipitation levels were provided by this station.

The Banjska River Basin was divided into two parts. The first, upstream part begins at the Prvonek Dam. Its surface area is 87,44 km^2 and its average altitude 1081.7 m. The second part of the river basin is the area downstream from the dam to the Vranjska Banja HS (see Fig. 2), which belongs to the monitoring system of the Serbian Hydrometeorological Service. The surface area of this portion of the river basin is 21.57 km^2, and its average altitude 701.1 km^2. The station was commissioned in 1963, when river stages were recorded by means of a staff gauge. A limnigraph was installed at this station in 1978. The Banjska River was impounded in 2005 and water has since then been extracted for drinking water supply. Previously, a water intake downstream from the present dam was used. The database of the Jaroslav Černi Institute for the Development of Water Resources contains data on discharges recorded by the Vranjska Banja HS from 1963 to 2008.

The hydrologic model was calibrated for the period from 1 October 1981 to 30 September 1983 at the Vranjska Banja HS. The simulated discharges (Q_s), measured discharges (Q_m) and precipitation levels (P) at the Kriva Feja station are shown in Fig.4. Snowmelt was simulated in two ways (Degree-Day and Temperature Index methods).

The Nash-Sutcliffe coefficient was used as the calibration criterion:

$$E = 1 - \frac{\sum_{i=1}^{n}(Q_{m,i} - Q_{s,i})^2}{\sum_{i=1}^{n}(Q_{m,i} - \overline{Q_m})^2}.$$

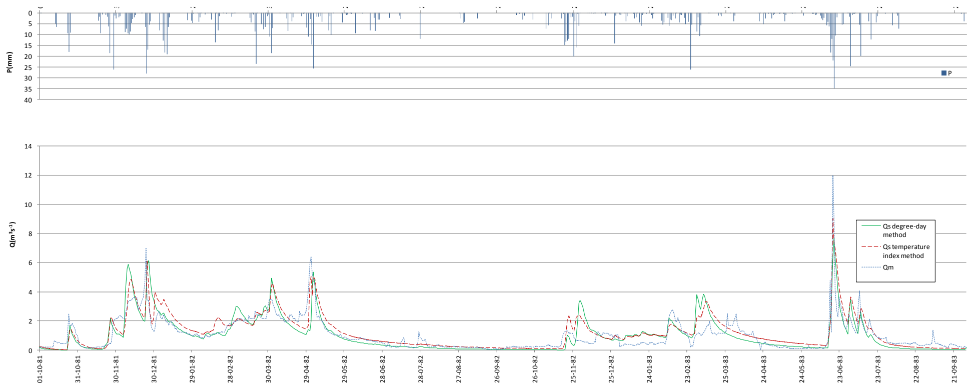

Fig. 4 Measured (Q_m) and simulated (Q_s) average daily discharges at the Vranjska Banja HS on the Banjska River for the calibration period from 1 October 1981 to 30 September 1983.

This coefficient amounts to E=0.720 for snowmelt applying the Temperature Index Method and E=0.588 according to the Degree-Day Method, which is a satisfactory assessment. Fig. 5 shows parts of the hydrologic cycle for the calibration period at the Vranjska Banja HS for snowmelt according to the Degree-Day (a) and the Temperature Index (b) method.

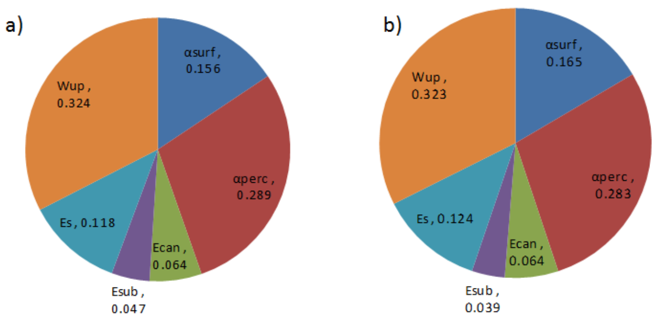

Fig. 5 Parts of the hydrologic cycle for the simulated period from 1 October 1981 to 30 September 1983 at the Vranjska Banja HS: Degree-Day Method (a) and Temperature Index Method (b).

The hydrologic model was then validated for the period from 1 October 1983 to 30 September 1985. Fig. 6 shows simulated discharges for the validation period, applying both snowmelt models.

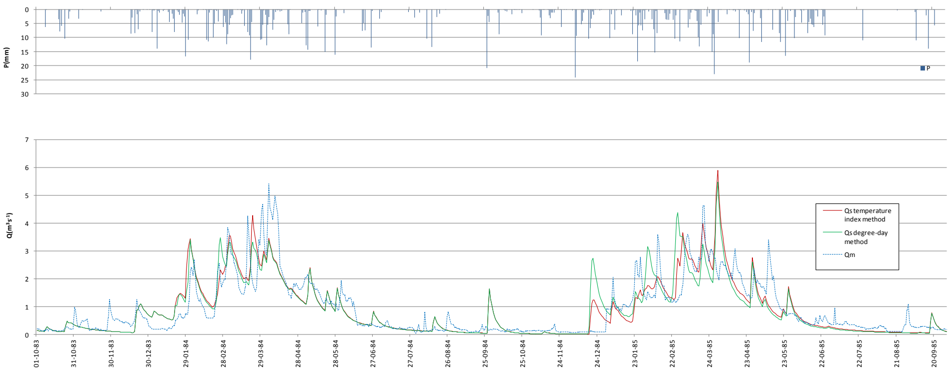

Fig. 6 Measured (Q_m) and simulated (Q_s) average daily discharges at the Vranjska Banja HS on the Banjska River for the validation period from 1 October 1983 to 30 September 1985.

5. Conclusion

A physically-based distributed hydrologic model was used in this research and applied to the Banjska River Basin. The Banjska River is a tributary of the Južna Morava. The river basin was represented by a grid of squares (*HRUs*), where the vertical water balance was assessed in each square through three reservoirs. Each *HRU* carried information about river basin morphology, vegetation and pedological and hydrogeological compositions of the soil. The effect of weather stations was defined by means of *Thiessen* polygons, including altitude correction.

The hydrologic model was partitioned into three reservoirs. The snow accumulation and snowmelt processes were defined in the snow reservoir applying the Degree-Day Method and the Temperature Index Method. Horizontal water movement to the Vranjska Banja Hydrological Station (P=109 km^2) was assumed to take place adhering to the linear reservoir principle.

The conclusion was that the hydrologic model based on the snowmelt temperature index exhibited a better match with measured data. Maximum discharges showed a greater similarity to recorded levels. This model yielded lower surface runoff and higher base runoff. Greater evaporation from the snow (sublimation) was noted, likely as a result of the greater depth of the snow cover and slower melting compared to the Degree-Day Method.

The Degree-Day method exhibited a better base runoff match. The snowmelt hydrograph branch was of a shorter duration due to more rapid snowmelt. The initial soil moisture levels were found to be lower prior to the onset of the dry period.

According to the Nash-Sutcliffe coefficient, the Temperature Index Method and the Degree-Day Method exhibited a satisfactory match with measured discharges. The Temperature Index Method is recommended for high-flow forecasts and the Degree-Day Method is proposed for monitoring of the decreasing branch of the hydrograph.

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HIDROLOŠKO MODELIRANJE SLIVA SA POSEBNIM OSVRTOM NA PROCESE U SNEŽNOM POKRIVAČU

U radu je korišćen fizički zasnovan hidrološki model sa rasprostranjenim parametrima. Slivno područje je diskretizovano mrežom kvadrata gde će svaki kvadrat ima podatke o morfološkim karakteristikama dela sliva, biljnom pokrivaču, pedološkom sastavu tla, hidrogeološkom sloju i drugo. Uticaj meteoroloških stanica definisan je Thiessen-ovim poligonima uz korišćenje korekcije po visini. Hidrološki model je kontinualan sa vremenskom diskretizacijom koja iznosi jedan dan. Hidrološki model je dekomponovan na tri rezervoara, pri čemu je rezervoar snega definisan metodom Stepen-Dan i metodom Temperaturnog Indeksa. Hidrološki model je primenjen na sliv Banjske Reke koja predstavlja pritoku Južne Morave.

Ključne reči: *Hidrološko modeliranje, akumulisanje snega, otapanje snega*