UNSTEADY FLOW ON RIVER BASIN SLOPE AND IN THE RIVER CHANNELS

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Contents

Summary

Surface flow, which includes the overland flow and the river channel flow, is an important part of the hydrological cycle and plays a key role in river runoff generation and water quality formation. Modeling of unsteady surface flow is one from most often used procedures in applied hydrology and in simulation of environmental processes. The choice of main factors and time-space scales for modeling of the surface flow is defined by the type of a given hydrological task, available information, and requirements to accuracy of calculations. As a result, the models of surface flow may have significant differences in representation of physical reality. This chapter presents a review of models of the surface flow and numerical procedures, which have been tested for application of these models for solving different problems of applied hydrology. The review begins from description of the two-dimensional hydrodynamic model of overland flow and then different hypotheses and assumptions (conceptions) are considered to allow simplifying the models. The one-dimensional channel flow models including the St. Venant equations and their simplifications (linear distributed models, the diffusion simplification, and the kinematic wave equations) are presented. The methods of numerical integration of unsteady surface flow equations are described. In many cases, the main goal of flood modeling is only transformation of the input hydrographs into the output hydrographs and the hydrological system can be considered

as a lumped linear or nonlinear hydrological systems. A part of the chapter is devoted to main assumptions and the procedures of application of widely used lumped linear time invariant and time-invariant models. Special attention is paid to determination of parameters of these models. Possibilities of constructing lumped nonlinear model based on the Volterra functional series are outlined.

1. Introduction

Modeling of unsteady surface runoff, which includes the overland flow and the river channel flow, is one of the most often used procedures in applied hydrology and in simulation of environmental processes. The form of hydrograph and the speed of flood movement depend mainly on the surface flow, and in many cases the prediction of flood characteristics can be limited by modeling of the surface flow (this procedure is often called flood routing).The choice of main factors and time-space scales in the construction of models of the surface flow is defined by the type of the given hydrological task, available information, and requirements of accuracy of calculations. The model can be developed for description of the surface flow as a component of the hydrological cycle or for only flood routing. As a result, the models of surface flow may have significant differences in representation of physical reality. Some models are based on strict description of hydrodynamic processes and using as model parameters that can be found by direct measurements or by laboratory experiments (these models are usually called “physically based”). Most routing models are based on introducing different hypotheses and assumptions (conceptions) to simplify the description of processes and decrease the number of parameters which can not be measured directly (these models are often called conceptual ones). In many cases, the main goal of routing models is only transformation of the input hydrographs into the output hydrographs and the hydrological system can be considered as a “black box”. Most parameters of conceptual and black box routing models and some parameters of the physically based routing models are found by adjusting parameter values of a model to obtain minimum difference between observed and calculated model output variables (this procedure is called calibration).

In this chapter, we tried to present the main assumptions, which commonly use in descriptions of unsteady flow on the surface of river basins and the routing models, which have appeared to be the best for application in the hydrological practice

2. The Two-Dimensional Models of Overland Flow

The two-dimensional continuous overland flow is observed on river slopes seldom and during short periods. Generally, the rainfall or snowmelt water excess appearing at the soil surface as sheet flow quickly reaches a temporary stream network. The structure of this network varies depending on the magnitude and the spatial distribution of the water excess, however the dominant direction of flow in the streamlets, their density and geometric characteristics are sufficiently stable and related to river slope topography and soil properties. It is also possible to assume that there is at least one streamlet in each minimum grid area chosen for runoff computing. Then, neglecting the process of water flow to the streamlet network, we can use for description of the streamlet network flow the same equations as for two-dimensional fictitious continuous overland flow.

The depth of this fictitious sheet flow will be significantly less than the real depth of flow in the streamlet system; however it is possible to preserve the discharges in each direction. Such an approach makes acceptable application of classic equations of hydrodynamics for describing water movement along river slopes.

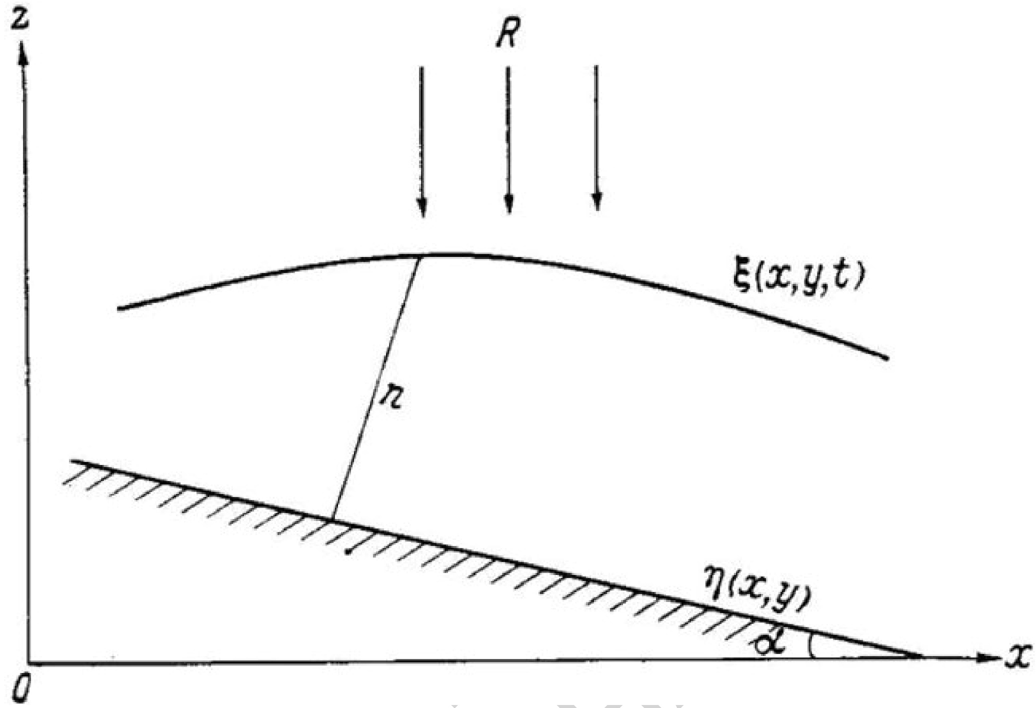

Figure1. Schematization of surface flow on the river slope

The equations of water flow along two-dimensional slope (Figure 1) can be presented as

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}+w\frac{\partial u}{\partial z}=-\frac{1}{\rho}\frac{\partial p}{\partial x}+\frac{1}{\rho}\frac{\partial \tau_x}{\partial z} \quad (1)$$

$$\frac{\partial v}{\partial t}+u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}+w\frac{\partial v}{\partial z}=-\frac{1}{\rho}\frac{\partial p}{\partial y}+\frac{1}{\rho}\frac{\partial \tau_y}{\partial z} \quad (2)$$

$$\frac{\partial w}{\partial t}+u\frac{\partial w}{\partial x}+v\frac{\partial w}{\partial y}+w\frac{\partial w}{\partial z}=-g-\frac{1}{\rho}\frac{\partial p}{\partial z} \quad (3)$$

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0 \quad (4)$$

where u, v and w are the velocity components in the x, y and z directions respectively (z is defined as being positive upward from an arbitrary datum with the ground), ρ is the density of water, p is the hydrodynamic pressure, g is the gravitational acceleration, τ_x and τ_y are the components of flow resistance stress in the x and y

directions, respectively.

This form of water flow equations can be obtained from the Navier-Stokes equations (for laminar flow) as well as for the Reynolds equations (for turbulent flow).

The first three equations represent conservation of momentum in the x, y and z directions, and the last equation represents mass conservation (or continuity).

At the upper boundary of flowing layer $z = \xi(x, y, t)$, we can impose the condition

$$w + R = \frac{\partial \xi}{\partial t} + u_0 \frac{\partial \xi}{\partial x} + v_0 \frac{\partial \xi}{\partial y} \qquad (5)$$

where R is the rainfall rate, u_0 and v_0 are horizontal components of the flow velocity at $\xi(x, y, t)$.

At the lower boundary $z = \eta(x, y)$, it is possible to assume $u = u_0 = 0$ and $\omega(\eta) = -I$ where I is the rate of infiltration into soil.

The horizontal scales of overland runoff are much more than flow depth change. Assuming the vertical velocity is negligibly small, we carry out the vertical averaging of system (1)-(4). The integration of (4) from z to ξ gives the hydrostatic law of pressure distribution (the shallow-water assumption)

$$p(z) = \rho R \lambda + P_a + \rho g(\xi - z),$$

where λ is the final vertical rainfall velocity, P_a is the atmospheric pressure.

Averaging of Eqs. (1), (2) and (4) results in

$$\frac{\partial}{\partial t}\int_\eta^\xi u\, dz + \frac{\partial}{\partial x}\int_\eta^\xi u^2\, dz + \frac{\partial}{\partial y}\int_\eta^\xi uv\, dz - u_0 R == -\frac{1}{\rho}\int_\eta^\xi \frac{\partial p}{\partial x}\, dz + \int_\eta^\xi \frac{\partial \tau_x}{\partial z}\, dz \qquad (6)$$

$$\frac{\partial}{\partial t}\int_\eta^\xi v\, dz + \frac{\partial}{\partial x}\int_\eta^\xi v^2\, dz + \frac{\partial}{\partial y}\int_\eta^\xi uv\, dz - v_0 R = -\frac{1}{\rho}\int_\eta^\xi \frac{\partial p}{\partial y}\, dz + \int_\eta^\xi \frac{\partial \tau_y}{\partial z}\, dz \qquad (7)$$

$$\frac{\partial \xi}{\partial t} + \frac{\partial}{\partial x}\int_\eta^\xi u\, dz + \frac{\partial}{\partial y}\int_\eta^\xi v\, dz = R - I \qquad (8)$$

In order to simplify this system, we introduce

$$q_x = \int_{\eta}^{\xi} u\, dz\,,\ \ q_y = \int_{\eta}^{\xi} v\, dz\,, \tag{9}$$

$$\int_{\eta}^{\xi} u^2\, dz \approx \alpha_x \bar{u}\, q_x\,,\ \int_{\eta}^{\xi} uv\, dz \approx \alpha_y \bar{v}\, q_x = \alpha_x \bar{u}\, q_y\,,\ \int_{\eta}^{\xi} v^2\, dz \approx \alpha_y \bar{v}\, q_y\,, \tag{10}$$

where $\bar{u} = \dfrac{q_x}{\xi - \eta}$, $\bar{v} = \dfrac{q_y}{\xi - \eta}$, α_x and α_y are the coefficients which take into account non-uniform vertical distribution of u and v respectively.

Assuming that the atmospheric pressure does not change over the river basin, i.e. $\dfrac{\partial P_{\mathrm{a}}}{\partial x} = 0$ and $\dfrac{\partial P_{\mathrm{a}}}{\partial y} = 0$ results in

$$\frac{1}{\rho}\int_{\eta}^{\xi} \frac{\partial P}{\partial x} dz = (\xi - \eta)\frac{\partial}{\partial x}(R\Lambda) + \frac{g}{2}\frac{\partial}{\partial x}(\xi - \eta)^2 + g(\xi - \eta)\frac{\partial h}{\partial x}. \tag{11}$$

$$\frac{1}{\rho}\int_{\eta}^{\xi} \frac{\partial P}{\partial y} dz = (\xi - \eta)\frac{\partial}{\partial x}(R\Lambda) + \frac{g}{2}\frac{\partial}{\partial y}(\xi - \eta)^2 + g(\xi - \eta)\frac{\partial h}{\partial y}. \tag{12}$$

Finally, it is necessary to choose a representation of resistance forces. In the general form these forces can be formally represented as

$$\int_{\eta}^{\xi} \frac{\partial \tau_x}{\partial z} dz = \frac{1}{\rho}\left(\tau_{\mathrm{s}x} - \tau_{\mathrm{b}x}\right) \tag{13}$$

$$\int_{\eta}^{\xi} \frac{\partial \tau_y}{\partial z} dz = \frac{1}{\rho}\left(\tau_{\mathrm{s}y} - \tau_{\mathrm{b}y}\right) \tag{14}$$

where $\tau_{\mathrm{s}x}$ and $\tau_{\mathrm{s}y}$ are the shear stresses caused by wind action at the water surface along direction x and y, respectively; $\tau_{\mathrm{b}x}$ and $\tau_{\mathrm{b}y}$ are the resistance stresses along direction x and y, respectively, The resistance stresses consist of bottom friction and lateral stresses that represent the combined effect of viscous and turbulent stresses and momentum transfers due to the vertical velocity distribution.

The conventional expressions for estimation of shear stress at the water surfaces are

$$\tau_{\mathrm{s}x} = k_{\mathrm{s}}\rho_{\mathrm{a}} w^2 \cos\varphi \tag{15}$$

$$\tau_{\mathrm{s}y} = k_{\mathrm{s}}\rho_{\mathrm{a}} w^2 \sin\varphi \tag{16}$$

where k_{s} is an empirical drag coefficient, ρ_{a} is the density of the air, w is the wind velocity, φ is the angle between the direction of the wind and the positive x-axis.

The usual assumption for bottom shear friction is that its magnitude is the same as one corresponding to steady uniform flow and that it acts in the direction of depth-average velocity.

For one-dimensional flow with the average cross-sectional velocity v, bottom friction can be represented as

$$\tau_{\mathrm{b}} = \frac{\rho g v^2}{C^2} \tag{17}$$

where C is the roughness coefficient for uniform flow in open channals used in the well-known Chezy equation

$$v = C[R_h S_0]^{0.5} \tag{18}$$

where R_{h} is the hydraulic radius equal to A/P_{h}, A is the cross-section area, P_{h} is the wetted perimeter, S_0 is the bottom channel slope. For broad channels the hydraulic radius R_{h} is approximately equal to the flow depth h.

If overland flow occurs in a very thin layer having large horizontal dimensions, the lateral stress may be neglected; however this stress may be important in small streams.

The contribution of viscous stresses in the depth-averaged lateral stress is typically quite small in comparison to the turbulent stress and may be essential only for laminar flow. The influence of the vertical velocity distribution can be neglected or be taking into account partially by choice of the coefficients α_x and α_y in (10).

Theory of turbulence gives a large number of options for parameterization of turbulent stresses; however lack of needed measurements allows for application only the simplest forms of such parameterizations. Among them is the oldest proposal for representation of the turbulent stresses that was formulated in 1877 by Boussinesq who assumed the turbulent stresses to be directly proportional to the mean-velocity gradients. The proportionally constant in this relation is called a turbulent-exchange coefficient or eddy viscosity and is analogous to the coefficient of molecular viscosity. The numerous theoretical formulae suggested for determination of eddy viscosity and eddy-viscosity terms in the momentum equations have found for modeling of natural water flow a limited application. In most cases, there is no information to separate the bottom stress and the turbulent stress. As a result, in many practical applications the effective coefficients used for representation of the bottom stress and determined with aid of calibration procedures may include also the turbulent stresses. Another possibility to account for the turbulent stresses is fitting of relevant computational viscosity when numerical methods are applied for the solution of the differential equations.

Thus, if the horizontal gradients of velocity are not too much, the resistance terms including the bottom and turbulence stress may be represented in the Eqs. (6-7) on the basis of Chezy equation with an effective value of the coefficient of roughness. The directional components τ_{bx} and τ_{by} are then given by

$$\tau_{bx} = \rho g \frac{u}{C^2 \left(u^2 + v^2\right)^{1/2}} \tag{19}$$

and

$$\tau_{by} = \rho g \frac{v}{C^2 \left(u^2 + v^2\right)^{1/2}} \tag{20}$$

Substituting these relations in the system (10-12), one can write this system in the following (conservative) form:

$$\frac{\partial \xi}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = R - I \tag{21}$$

$$\begin{aligned} &\frac{\partial q_x}{\partial t} + \frac{\partial}{\partial x}\left(\alpha_x \bar{u}\, q_x\right) + \frac{\partial}{\partial y}\left(\alpha_x \bar{u}\, q_y\right) - u_0 R = \\ &\quad = -\frac{g}{2}\frac{\partial (\xi-\eta)^2}{\partial x} - (\xi-\eta)\frac{\partial}{\partial x}(R\Lambda) - g(\xi-\eta)\frac{\partial \eta}{\partial x} - g\frac{u\left(u^2+v^2\right)^{1/2}}{c^2}; \end{aligned} \tag{22}$$

$$\begin{aligned} &\frac{\partial q_y}{\partial t} + \frac{\partial}{\partial x}\left(\alpha_y \bar{v}\, q_x\right) + \frac{\partial}{\partial y}\left(\alpha_y \bar{v}\, q_y\right) - v_0 R = \\ &\quad = -\frac{g}{2}\frac{\partial (\xi-\eta)^2}{\partial y} - (\xi-\eta)\frac{\partial}{\partial y}(R\Lambda) - g(\xi-\eta)\frac{\partial \eta}{\partial y} - g\frac{v\left(u^2+v^2\right)^{1/2}}{C^2} \end{aligned} \tag{23}$$

Taking into account the fact that $\frac{\partial q_x}{\partial t} = u\frac{\partial}{\partial t}(\xi-\eta) + \frac{\partial \bar{u}}{\partial t}(\xi-\eta)$ and substituting $\frac{\partial \xi}{\partial t}$ from (21) in the Eq. (21), we obtain

$$\begin{aligned} &(\xi-\eta)\frac{\partial \bar{u}}{\partial t} + u(\xi-\eta)\frac{\partial \bar{u}}{\partial x} + \bar{v}(\xi-\eta)\frac{\partial \bar{u}}{\partial y} + g(\xi-\eta)\frac{\partial}{\partial x} + \left(\bar{u} - u_0\right)R - \bar{u}\, I = \\ &\quad = -g(\xi-\eta)\frac{\partial h}{\partial x} - g\frac{u\left(u^2+v^2\right)^{1/2}}{C^2} - (\xi-\eta)\frac{\partial}{\partial x}(R\Lambda) \end{aligned} \tag{24}$$

Assuming that most part of bottom surface has small slope and dividing both momentum equations by $h = (\xi - \eta)$, we obtain the two-dimensional equations of overland flow in the conventional (non-conservative) form:

$$\frac{\partial \xi}{\partial t}+\frac{\partial q_x}{\partial x}+\frac{\partial q_y}{\partial y}=R-I \tag{25}$$

$$\frac{\partial \bar{u}}{\partial t}+\bar{u}\frac{\partial \bar{u}}{\partial x}+\bar{v}\frac{\partial \bar{u}}{\partial y}+g\frac{\partial h}{\partial x}+\frac{(u-u_0)R-\bar{u}I}{h}= ; \\ =-g\frac{\partial \eta}{\partial x}-\frac{u\left(u^2+v^2\right)^{1/2}}{C^2 h}-\frac{\partial (R\Lambda)}{\partial x} \tag{26}$$

$$\frac{\partial \bar{v}}{\partial t}+\bar{u}\frac{\partial \bar{v}}{\partial x}+\bar{v}\frac{\partial \bar{v}}{\partial y}+g\frac{\partial h}{\partial y}+\frac{(v-v_0)R-\bar{v}I}{h}= \\ =-g\frac{\partial \eta}{\partial y}-\frac{v\left(u^2+v^2\right)^{1/2}}{C^2 h}-\frac{\partial (R\Lambda)}{\partial y} \tag{27}$$

In order to compare the orders of the terms of Eqs. (25-27) on the basis of available experimental data, one may assign the reference scales of overland flow that are presented in Table 1.

Flow type	**Slope length, *l*, m**	***h*, m**	***u*, m/s**	***v*, m/s**	***w*, m/s**	***t*, s**	$\frac{\partial \eta}{\partial x}$	***R*, m/s**	Λ**, m/s**	***I*, m/s**	C^2**, m/s²**
Sheet	10^{2}	10^{-3}	10^{-2}	10^{-2}	10^{-5}	10^{3}	10^{-4}	10^{-5}	1	10^{-5}	10^{-3}
Streamlet	10^{3}	10^{-2}	10^{-1}	10^{-1}	10^{-4}	10^{4}	10^{-3}	10^{-5}	1	10^{-5}	10^{-3}

Table 1. The reference scales of overland flow.

The estimated orders of the terms are given in Table 2.

Flow type	$\frac{\partial \bar{u}}{\partial t}$	$\bar{u}\frac{\partial \bar{u}}{\partial x}$	$\bar{u}\frac{\partial \bar{u}}{\partial y}$	$\frac{\bar{u}(R-I)}{h}$	$g\frac{\partial h}{\partial x}$	$\frac{\partial}{\partial x}(R\Lambda)$	$g\frac{\bar{u}\sqrt{\bar{u}^2+v^2}}{C^2 h}$
Sheet	10^{-5}	10^{-6}	10^{-6}	10^{-4}	10^{-3}	10^{-7}	10^{-3}
Streamlet	10^{-5}	10^{-5}	10^{-5}	10^{-4}	10^{-2}	10^{-8}	10^{-2}

Table 2. The orders of the terms of momentum Eq. (26).

As can be seen from Table 2, the term

$$\frac{\partial}{\partial x}(R\Lambda)$$

is negligibly small for both sheet and streamlet flow. The inertial-force terms are two-three orders less than the gravitational and resistance terms, and these terms can also be commonly neglected, especially in the cases of sheet flow. However, at very small slope and relatively large flow velocities, the influence of inertial-force terms can be significant.

The system (25-27) belongs to the symmetric hyperbolic partial differential equations and according to the theory of such systems it can be transformed into equations that describe the surfaces or lines in the phase space (x, y, t) along which some functions of independent variables are invariant (Abbott, M.B. (1979)). These surfaces are called the characteristics and the corresponding functions are called the Riemann invariants. If any disturbances appear at a point of the solution domain, they propagate in the phase space along characteristics and the characteristics are often called the surfaces of disturbance distribution. The number of real characteristics crossing each point of the solution existence domain is equal to the number of independent variables. Consequently, the values of unknown functions at each point of the solution domain are determined by three characteristics. The number of characteristics crossing boundaries of the solution domain can be different and depends on the solution. In order to determine the Riemann invariants, it is necessary to assign additional conditions at the boundary (boundary conditions). Thus, the number of needed boundary conditions depends on the number of real characteristics crossing the boundaries. It is difficult to establish this number of needed conditions for an arbitrary solution domain if the solution behavior is unknown. However, the experience of solution of two-dimensional fluid flow equations in geophysics has shown that it is possible to use the following rule: for subcritical flow (i.e. at the Froude number $Fr = u^2/gh < 1$) at the boundary where there is inflow of liquid it is necessary to specify two boundary conditions; at the boundary where there is outflow of liquid it is necessary to specify one boundary condition.

Taking into account this rule and experience of solution of hydrological tasks, it is possible to recommend for solving (25-27) to specify the following boundary conditions: a) at boundary of river basin where there is not inflow the depth of flow and the tangential component of velocity can be assigned to be equal to zero; b) at the boundary where there is outflow the different relations between depths and velocities can be assigned.

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Biographical Sketch

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