MATHEMATICAL MODELING OF FLOW IN WATERSHEDS AND RIVERS

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Contents

Summary

New flow modeling techniques currently being developed are either based on *ANN*, *FL*, or *GA*, or are part deterministic and part stochastic. Watersheds and rivers are inherently spatial and complex, and our understanding of flow in these systems is less than complete. Many of the flow systems are either fully stochastic or part-stochastic and part-deterministic. Their stochastic nature can be attributed to randomness in one or more of the following components that constitute them: (1) system structure (geometry), (2) system dynamics, (3) forcing functions (sources and sinks), and (4) initial and boundary conditions. As a result, a stochastic description of these systems is needed, and the statistical techniques are available which enable development of such a description. Stochastic techniques are based on either point estimation methods or probability distribution functions. Copulas have tremendous potential in describing dependence between flow variables.

1. Introduction

Atmosphere is the source of water that flows overland and in channels, even though the amount of water stored in the atmosphere as water vapor is small, as compared with that in oceans and seas, polar icecaps, lakes and streams, and ground water. The atmospheric water falls on the land surface, streams, lakes, ponds, and seas and oceans as precipitation. Part of this precipitation returns to the atmosphere through evaporation as water vapor, and part of it may either run off or get stored. The remainder fills in the depressions on the ground, meets the infiltrative demand of the soil, and runs off the ground to form stream flow. The infiltrated water percolates down and recharges groundwater and may eventually become stream flow. The final destination of all streams is ocean, so streamflows finally reach seas and oceans. This cyclic movement of water from the atmosphere through precipitation to the land, through stream flow to the ocean, and through evaporation and evapotranspiration back to the atmosphere is called the hydrologic or water cycle. The movement of water, of course, follows devious paths and occurs in different directions. The flow of water in watersheds and rivers represents one component of this cycle, and constitutes the subject matter of this chapter.

2. Flow in Watersheds and Channels

A watershed is comprised of land areas and channels and may have lakes, ponds or other water bodies, depending on its size. The flow of water on land areas occurs not only over the surface but also below it in the unsaturated zone immediately below and further below in the saturated flow. The water over the land surface flows as overland flow; it occurs in both directions, longitudinal and transverse but its predominant direction is longitudinal. Thus, although overland is two-dimensional, its one-dimensional approximation is acceptable for most cases of practical interest. The flow in the unsaturated zone, called unsaturated flow, occurs predominantly vertically downward. The flow in reality is three dimensional but its one-dimensional or at most two-dimensional approximation suffices for most cases of interest. Of course the predominant direction of unsaturated soil moisture flow may change as the degree of saturation changes. When the soil becomes saturated the predominant flow direction changes and becomes longitudinal. The flow in the saturated zone, groundwater flow or baseflow, occurs principally longitudinally. In this case also, the flow is three-dimensional but its two dimensional or even one dimensional approximation will be adequate for practical purposes. When water returns to the atmosphere, it moves upward and the flow is three-dimensional. However, the principal direction is vertically upward. The return of water to the atmosphere as vapor will not be included in this chapter.

The flow of water in channels, streams and rivers occurs primarily and predominantly in the longitudinal direction. Locally, the flow may occur in the transverse direction as well as in the vertical direction. Theoretically the flow is three dimensional but its one-dimensional approximation is adequate. The term channel is used in a broad sense and includes rivers, streams, bayous, brooks, creeks, canals, sewers, partially flowing pipes and tunnels, gutters, borders, and furrows.

It is clear from the above discussion that the movement of water in watersheds and rivers occurs in virtually all directions and is three-dimensional. The discussion in this chapter will be confined to one dimensional flow in the predominant-longitudinal-direction.

There is a multitude of flow modeling techniques which can be broadly classified as hydrologic and hydraulic. Hydrologic modeling is based on a spatially lumped form of the continuity equation, often called water budget or balance, and a flux relation expressing storage as a function of inflow and outflow (Singh, 1988). Since coupling of these two equations leads to a first order ordinary differential equation, only an initial condition is needed to solve this equation. This equation does not explicitly involve any spatial variability and expresses the flow variable as a function of only time.

Hydraulic modeling is based on the St. Venant equations or simplifications thereof. A vast amount of literature dealing with applications of these equations or their simplifications to flow modeling is available (Singh, 1996; ASCE, 1996). In a watershed there usually is a network of overland flow planes, river channels and tributaries, that is, each river may have a number of tributaries. For purposes of applying these equations, a given river may be divided into a number of reaches. The hydraulic equations are applied to each reach having its own drainage area and the system of equations corresponding to all the reaches are solved simultaneously. When the full St. Venant equations are applied, the computational demands may be formidable and the solution may be inefficient and may incur a large accumulated error. This may explain the reason for the increasing popularity of simplified hydraulic models. These simplified models include kinematic wave, diffusion wave, gravity wave, and quasi-steady state. Linearized forms of the St. Venant equations are also popular for flow modeling (Dooge, 1980).

For most problems of practical interest dealing with flow movement, analytical solutions for either full St. Venant equations or their simplified forms are not tractable and numerical solutions are therefore employed. Numerical methods to obtain solutions can be classified as: explicit finite difference, implicit finite difference, finite element, and boundary fitted coordinate (Singh, 1996). In each class there are many different types of methods. Different classes of methods are useful for different conditions.

3. Governing Equations

The laws that govern the movement of water over and below the ground or in rivers are the conservation of mass, momentum, and energy. For the movement of water in unsaturated and saturated zones below the ground, the momentum plays a relatively minor role and hence its conservation is not important. The conservation of energy is expressed by an appropriate flux law. For surface flow, the conservation of mass is expressed as a continuity equation, and that of momentum as an equation of motion. Depending on the flow conditions, these equations are expressed in a variety of forms, as will be clear from the following discussion.

3.1. Surface Flow

For simplicity, only the one-dimensional form of the governing equations using a control volume is given here. The continuity equation can be expressed as

$$\frac{\partial A}{\partial t}+\frac{\partial Q}{\partial x}=q(x,t)-i(x,t)-e(x,t) \tag{1}$$

the momentum equation as

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+g\frac{\partial h}{\partial x}=g(S_0-S_{\mathrm{f}})-\frac{(q-i)(u-v)}{A} \tag{2}$$

and the energy equation as

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+g\frac{\partial h}{\partial x}=g(S_0-S_{\mathrm{f}})+\frac{u-v(v/u)}{2A}q \tag{3}$$

where A is the flow cross-sectional area, Q is the discharge (volumetric rate $=u.A$), u is the average flow velocity, h is the depth of flow, S_0 is the bed slope, S_{f} is the frictional slope, q is the lateral inflow per unit length of flow, i is the infiltration per unit length, e is the evaporation rate and other abstractions per unit length, v is the velocity of lateral inflow in the longitudinal direction, x is the distance in the longitudinal direction, and t is time. Except for the term expressing the influence of lateral inflow or outflow, Eqs. (2) and (3) are equivalent. Therefore, only Eq. (2) will henceforth be referred to.

Equation (1), in conjunction with either Eq. (2) or (3), can be employed to model surface flows on plains and/or in channels. Two popular approximations of Eq. (2) are the diffusion-wave and kinematic-wave approximations (Lighthill and Whitham, 1955;

Dooge, 1973; Singh, 1996), which can be expressed, respectively, as follows:

$$\frac{\partial h}{\partial x} = S_0 - S_{\mathrm{f}} \tag{4}$$

$$S_0 = S_{\mathrm{f}} \tag{5}$$

With use of a uniform flow formula such as Manning's or Chezy's, S_{f} can be expressed as

$$S_{\mathrm{f}} = \beta \frac{u^2}{R^a} \tag{6}$$

where $\beta = 1/C^2$ and $a = 1$ for Chezy's equation; $\beta = n_{\mathrm{m}}^2$ and $a = 4/3$ for Manning's equation; C is Chezy's roughness coefficient, n_{m} is Manning's roughness factor, and R is the hydraulic radius (= A/P, P = wetted perimeter).

Substitution of Eq. (6) into Eq. (5) and recognizing the unique relation between R and h leads to

$$u = \alpha h^m, \quad m > 0 \quad \text{or} \quad Q = \alpha h^n, \quad n = m + 1 \tag{7}$$

where m = 0.5 and α = $C(S_0)^{0.5}$ for Chezy's equation, and m = $2/3$ and α = $(S_0)^{0.5}/n_{\mathrm{m}}$ for Manning's equation. Equation (7) is normally applied to wide rectangular cross-sections. Otherwise, Q can also be expressed in terms of A in place of h. The kinematic-wave approximation hypothesizes a unique relationship between the flux (average velocity), concentration (depth), and position. Thus, this approximation can also be expressed in forms different from Eq. (7), as shown by Beven (1979).

Woolhiser and Liggett (1967) showed that the kinematic wave approximation would be adequate if the kinematic wave number K, was greater than or equal to 20. K can be expressed as $K = (S_0 L_0)/(h_0 F_0^2)$ where h_0 is the normal depth, L_0 is the flow plane length, F_0 is the Froude number for normal flow = $u_0/(gh_0)^{0.5}$, and u_0 is the normal velocity, This criterion is satisfied if the channel has a moderate slope and the flow is unsteady gradually varying and has little backwater effects (Ponce and Simons, 1977; Ponce et al., 1978; Hunt, 1984; Singh, 1996). The diffusion wave or non-inertia wave approximations are an improvement over the kinematic wave approximation because they are capable of accommodating backwater effects (Akan and Yen, 1977; Hager and Hager, 1985; Dooge and Napiorkowski, 1987; Singh, 1996). Likewise, gravity wave approximations perform well when inertial effects dominate over slope terms (Singh, 1996). These simplified approximations are adequate in most cases of practical interest (Yen, 1979, 1982; Marsalek et al., 1996; Singh, 1996).

If the control volume is extended to the scale of a watershed or a channel segment, then the flow variables are lumped or integrated in space and only their temporal variability is retained. Thus, integration of Eq. (1) in space leads to

$$\frac{dS}{dt} = Q - I(t) - f(t) - E(t) \tag{8}$$

where

$$S = \int_{x_1}^{x_2} A\,dx, \quad Q = Q(x_2, t), \quad I = Q(x_1, t) + \int_{x_1}^{x_2} q\,dx,$$
$$f = \int_{x_1}^{x_2} i\,dx, \quad E = \int_{x_1}^{x_2} e\,dx$$

where S is storage or volume, and Q is discharge as volumetric rate. Equation (8) is a volume balance or water budget equation with two unknowns, S and Q. Its solution requires another equation relating S to Q, I, and/or other variables. A very general relation between S and I and Q is (Singh, 1988):

$$S = \sum_{j=0}^{M} a(Q,I)\frac{d^jQ}{dt^j} + \sum_{l=0}^{N} b(Q,I)\frac{d^lI}{dt^l} \tag{9}$$

where a and b are coefficients, and M and N are some integers. A special case, involving one of the most frequently used relations in hydrology, is $S = S(Q)$:

$$S = KQ, \quad S = kQ^{\beta} \tag{10}$$

where K is the storage parameter (lag time for $\beta = 1$), and k and β are parameters. Eq. (10) can be derived from the momentum equation. As an example, consider Eq. (7) with $n = 1$. By multiplying both sides by $dx \sim x_2 - x_1$ and recalling that $S = dx.h.1$ and Q is volumetric flow rate, Eq. (10) results immediately.

3.1. Unsaturated Flow

In the unsaturated zone below the land surface, part of the pore space is occupied by air, so the degree of saturation is to be taken into account when dealing with unsaturated flow. The moisture content θ in the medium (volume of water per unit volume of porous medium) is a function of the capillary pressure $\psi < 0$, and likewise is the medium's hydraulic conductivity $K(\psi)$. The basic governing equations for unsaturated flow are the continuity equation and a flux law given by Darcy's equation in lieu of the momentum equation or energy equation. This flux law can also be derived from energy conservation considerations. The three-dimensional continuity equation, under the assumption of incompressible water, can be written as

$$\frac{\partial q_x}{\partial x}+\frac{\partial q_y}{\partial y}+\frac{\partial q_z}{\partial z}=-\frac{\partial \theta}{\partial t} \tag{11}$$

and Darcy's equation as

$$q_s=-K_s(\psi)\frac{\partial h}{\partial s}, \quad s=x,y,z; \quad \vec{q}=\{q_x,q_y,q_z\} \tag{12}$$

where h is the hydraulic head and q_s is the flux in the s direction. Substituting Eq. (12) into Eq. (11) and recalling that $h = \psi + z$, one gets

$$\begin{aligned}&\frac{\partial}{\partial x}\left(K_x(\psi)\frac{\partial \psi}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y(\psi)\frac{\partial \psi}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z(\psi)\frac{\partial \psi}{\partial z}+K_z(\psi)\right)\\&=C(\psi)\frac{\partial \psi}{\partial t},\\&\quad C(\psi)=\frac{\partial \theta}{\partial \psi}\end{aligned} \tag{13}$$

where $C(\psi)$ is the specific moisture capacity. This is the well-known Richards equation (Richards, 1931). Based on simplifications of porous media properties (anisotropy and heterogeneity) and the nature of flow, a number of simpler versions can be derived (Singh, 1997b).

A popular approximation is the kinematic wave approximation which assumes a unique relation between flux and soil moisture content as:

$$q_s=K_s(\theta), \quad s=x,y,z; \quad \vec{q}=\{q_x,q_y,q_z\} \tag{14}$$

On the other hand, if the control volume is extended to a soil element, then spatially lumped equations can be derived. For example, Eq. (11) can be integrated over space and expressed in the form of a water balance equation as

$$\frac{dS(t)}{dt}=f_s(t)-f(t) \tag{15}$$

where $S(t)$ is the water stored in the soil element, $f_s(t)$ is the seepage rate from the element, and $f(t)$ is the infiltration rate. The seepage rate is analogous to steady infiltration rate and is a function of soil characteristics. For most soils, this rate has been tabulated and is given in standard hydrology textbooks. If the initial storage space available in the element is S_0, then the change in the water storage space at any time t is

$$W(t)=S_0-S(t)=\int_0^t [\,f(t)-f_s(t)\,dt \tag{16}$$

which is an integral expression of continuity.

Another relation between $f(t)$ and $S(t)$ in lieu of Eq. (12), proposed by Singh and Yu (1990), can be expressed as

$$f(t) = f_s(t) + \frac{a[S(t)]^m}{[S_0 - S(t)]^n} \tag{17}$$

where a, m, and n are positive real constants and are determined empirically. Equation (17) is a generalized flux relation for infiltration.

3.3. Saturated Flow

The governing equations for saturated flow are the continuity equation and the flux law specified by Darcy's equation. A three-dimensional form of continuity equation for incompressible flow is

$$\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z} = -S_s \frac{\partial h}{\partial t} \tag{18}$$

where S_s is the specific storage for confined formations, or specific yield divided by the saturated thickness for unconfined formations. Darcy's equation can be written as

$$q_s = -K_s \frac{\partial h}{\partial s}, \quad s = x, y, z; \quad \vec{q} = \{q_x, q_y, q_z\} \tag{19}$$

where K_s is the saturated hydraulic conductivity in the s direction. Substitution of Eq. (19) into Eq. (18) gives the general flow equation, which specializes-depending on the simplifications of porous media properties and the nature of flow-into a number of equations, such as the Laplace equation, the diffusion equation, the Theis equation, the Poisson equation, the Boussinesq equation, and so on (Singh, 1997b).

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