Principles of Hydrology

Soil Moisture

The unsaturated zone or vadose zone is referred to a region in which the pore spaces of the rock or soil may partly filled with air and water. The unsaturated zone consists of soil water zone, intermediate zone, and capillary fringe.

Soil moisture is water held in soils and rocks in the unsaturated zone.

After a rather rapid decrease from saturation, the moisture content in the intermediate belt may remain fairly constant This constant moisture content in the intermediate zone is Field capacity

Saturated value - in the capillary fringe.

Capillary-Pressure Head and the Moisture Characteristic

The relationship between the external suction applied to a rock and the amount of water per bulk volume (the moisture content) that the rock retains against that de-watering suction is called the moisture characteristic.

Darcy=s Law

Water flowing through soil and ground water is a part of the hydrologic cycle that lumped the ocean, atmosphere, and land areas. As we realize, the drinking water obtained from ground water can be easily contaminated through the leakage of contaminants in the soil from ground surface by various means. Understanding of water flow and contaminant transport along the pathways in the soil and ground water is crucial to the contamination prevention and remediation.

$$\frac{Q}{A} = q = -K(\theta)\frac{\partial h}{\partial z}$$

Water flow through soil as a steady flow may be described with Darcy=s equation: where Q is the volumetric flow rate (cm^3/day), A is the cross-sectional area (cm^2), q is the volumeteric flow rate per unit surface area, K(è) is the hydraulic conductivity (cm/day), è is the volumetric water content (cm^3/cm^3), h is the hydraulic head, and z is distance (cm). The specific discharge q in the saturated zone or ground water is a superficial or apparent velocity because water only moves through the pore openings making up the surface area (Domenico and Schwartz, 1990). The more realistic velocity (v) is a volumetric flow rate per unit area of connected pore space: $v = q / n_e$ (n_e - effective porosity).

There are thousands of measurements of hydraulic conductivity obtained in the field and laboratory over years. The values of hydraulic conductivity can range over 12 orders of magnitude from $1x10^{-8}$ cm/day in unfractured igneous and metamorphic rocks to tens or thousands of meter per day in gravels and some karstic or reef limestones (Davis, 1969). Various direct and indirect field and laboratory methods are available for measuring hydraulic conductivity. In unsaturated soils, the hydraulic conductivity is a function of saturated hydraulic conductivity and soil water content. The soil water content can be measured by simply drying samples, neutron probe, and time domain reflectometry method (TDR). The hydraulic conductivity and soil-water metric potential in schemes for evaluating hydraulic parameters are expressed as functions of the normalized volumetric water content (S_e), which is defined as

$$S_e = \frac{q - q_r}{q_s - q_r}$$

where $è_s$ is the soil water content at saturation, and $è_r$ is the residual soil water content. Various schemes are available for parametering soil hydraulic properties (Clapp and Hornberger, 1978; Van Genuchten, 1980; Cosby et al., 1984). The following equations have been widely used in various soil and hydrological applications. A simple equation by Clapp and Hornberger, 1978 and Cosby et al., 1984 can be used for relating hydraulic conductivity to soil water content:

$$K(q) = K_s\,(S_e)^{3+1/l}$$

where K_s is the saturated hydraulic conductivity (cm/day), and ë is a pore size parameter. A more versatile equation relating the hydraulic conductivity to soil water content by Rawls and Brakensiek (1985) can be described as:

$$K(q) = K_s\,(S_e)^{1/2}\,[1-(1-S_e^{1/m})^m]^2$$

where m = ë / (ë+1). Choice of the numerical values for these parameters is critical for correctly calculating soil water content.

The following equation can be used to describe the transient flow in the vertical direction:

$$\frac{\partial q}{\partial t} = -\frac{\partial q}{\partial z}$$

Combining Equation 5 with Darcy=s equation (Equation 3) yields the following equation (or Richards= Equation) for describing the rate of change of volumetric water content (Swartzendruber, 1969):

$$-\frac{\partial q}{\partial z} = \frac{\partial}{\partial z}[K \frac{\partial y}{\partial \boldsymbol{q}} \frac{\partial \boldsymbol{q}}{\partial z}] + \frac{\partial K}{\partial \boldsymbol{q}} \frac{\partial \boldsymbol{q}}{\partial z}$$

where ø is hydraulic potential. Analytic solutions are available for Equation 6 for some well-defined conditions. In general, the Richards= equation is solved numerically for the solution of soil moisture flow flux.

Infiltration and runoff

Horton=s Equation

Several rainfall-runoff generating processes have been recognized over the years (Dunne, 1978; Freeze, 1980; Beven, 1989). The transformation of precipitation into surface runoff is controlled by the independent interaction of many spatially variable processes. Horton runoff (Horton, 1933) and Dunne runoff (Dunne and Black, 1970) are perhaps the two most important conceptual models for surface runoff. Horton runoff is considered the excess of precipitation intensity over soil infiltration rate at a point (Freeze, 1974).:

$$f(t) = f_c + (f_0 - f_c)e^{-kt}$$

where f(t) is the infiltration at time t (cm/hr), f_0 is the initial infiltration rate (cm/hr), f_c is the constant infiltration rate (cm/hr), and k is a decay constant.

Philip=s Equation

Philip (1957) solved Richards= equation under less restrictive conditions by relating conductivity and diffusivity to the soil moisture content. The cumulative infiltration F can expressed

$$F = S\, t^{1/2} + Kt$$

where S is sorptivity. The infiltration rate at time t can be obtained by differentiating the above equation

$$f(t) = \frac{1}{2} S t^{-1/2} + K$$

Green-Ampt method

Green and Ampt (1911) developed approximate solutions of Richards= equation for infiltration calculation.

1, if rainfall intensity (i) is <= K_s, then f = i
2. If I > K_s, then f = I until F = it_s = F_s

$$F_s = \frac{[(q_s - q_i) y_f]}{[1 - i/K_s]} = \frac{M_d y_f}{[1 - i/K_s]}$$

where è$_s$ is the saturated moisture content, è$_i$ is the initial moisture content, ø$_f$ is tension or suction, and M_d is the initial moisture deficit.

$$f = K_s [1 - \frac{M_d y_f}{F}] \quad for\ i > K_s$$

3. Following surface saturation,
in the case of i = K_s, f=i.

Example, ...

Saturation runoff

Dunne runoff occurs when precipitation falls over a saturated area or impermeable surface. This flow is partially responsible for the rapid streamflow response associated with the expansion of saturated areas (or referred as to variable source areas) along streams or valleys during a storm event (Dunne and Black, 1970). Various models including TOPMODEL by Beven and Kirkby (1979) were developed to study various source areas (Troendle, 1985).

SCS curve number method

Rainfall excess or surface runoff can be determined using a number of rainfall-runoff schemes. A commonly-used method is the curve number approach developed by the Soil Conservation Service (USDA-SCS, 1972). The curve number method can be easily applied to large regions and implemented using a GIS, making this approach suitable for the evaluation of spatial variability at various resolutions. The accumulated runoff or rainfall excess Q(t) at time t can be expressed as

$$Q(t)=\frac{(P(t)-I_a)^2}{(P(t)-I_a+S)}$$

where P(t) is the accumulated precipitation, and I_a is the initial abstraction before ponding (i.e., the maximum amount of rainfall which can be retained on the surface before runoff occurs). The potential maximum retention S is defined as

$$S=\frac{1000}{CN}-10$$

where CN is the curve number.

Without considering interception and evapotranspiration, the cumulative infiltration F(t) after time t can be expressed as the difference between the cumulative precipitation and the sum of the cumulative runoff and the initial abstraction

$$F(t)=P(t)-Q(t)-I_a$$

The change dQ in accumulated runoff during a time interval dt may be defined by differentiating Equation 1 with respect to t and rearranging terms. This yields

$$dQ=[1-\frac{S^2}{(P(t)-I_a+S)^2}]\,dP$$

where dP is the incremental precipitation during dt. The initial abstraction I_a was approximated in this study by the commonly used value 0.2S.

Streamflow

A streamflow or discharge hydrograph is a graph or table showing the flow rate as a function of time at a given location on the stream. The streamflow consists of surface runoff, baseflow, and return flow.

Streamflow separation.
Example, ...