TRAINING COURSE
ON

HYDROLOGICAL MODELING AND GIS

(MAY 26 TO JUNE 06, 2014)

FOR

UNFAO & Ministry of Energy and Water, Afghanistan

LECTURE NOTE
ON

INTRODUCTION TO HYDROLOGIC MODELING

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INTRODUCTION TO HYDROLOGIC MODELING

INTRODUCTION

Rainfall-runoff modeling is an important aspect of hydrologic analysis and design. Choice of an appropriate approach to modeling of rainfall to runoff transformation process in a basin is influenced by various factors which include (i) typical features of the system; (ii) objectives of the study; (iii) degree of realism; and (iv) availability of data and resources; and (v) time scale of analysis.

A study of a river basin's hydrology and consequently, the design of suitable study approach, by its very nature is quite extensive. While the design of suitable hydrologic methodologies is an important consideration equally critical is the manner of application of these methodologies. It is impossible to build exact scale models of the hydro-climatological systems on which one could perform experiments to understand the nature of its operations on rainfall and its eventual transformation into runoff. The mathematical modeling approach is an alternative path. Evolution of the art of mathematical modeling of basin hydrology, often used for the purpose of runoff simulation, has followed several courses. We first explain here the rainfall-runoff process.

RAINFALL-RUNOFF PROCESS

The antecedent conditions as well as the volume and intensity of precipitation will be important in governing the processes by which a catchment responds to the input precipitation and the proportion of the input volume that appears as output hydrograph. Unless the stream is ephemeral, there will always be a response to the precipitation falling directly on the channel and nearby areas. This factor might be important contributor to the hydrograph in catchments and storms with low runoff coefficients. Even in ephemeral streams, surface flow will often start first in the stream channels. The extent of the channel network will generally expand into headwater areas as a storm progresses and will be greater during wet seasons than dry.

Rainfall and snowmelt inputs are not spatially uniform, but can show rapid changes in intensity and volume over relatively short distances, particularly in convective events. The variability at ground level, after the pattern of intensities has been affected by the vegetation canopy, may be even greater. Some of the rainfall will fall directly to the ground as direct throughfall. Some of the rainfall will be intercepted and evaporated from the canopy back to the atmosphere. Some evaporation of intercepted water may occur even during events, especially from rough canopies, under windy conditions, when the air is not saturated with vapour. The remaining rainfall will drip from the vegetation canopy as throughfall or run down the branches, trunks and stems as stemflow. The latter process may be important since, for some canopies, 10 percent or more of the incident rainfall may reach the ground as stemflow, resulting in local concentrations of water at much higher intensities than the incident rainfall.

Snowmelt rates will vary with elevation and aspect in that they are affected by the air temperature and radiation inputs to the snowpack. The water equivalent of the snowpack can vary dramatically in space, particularly due to the effects of wind drifting during snow events and due to the topography and vegetation cover.

Once the rain or snowmelt water has reached the ground, it will start to infiltrate the soil surface, except on impermeable areas of bare rock, on areas of completely frozen soil, or some artificial surfaces where surface runoff will start almost immediately. The rate and amount of infiltration will be limited by the local ground level rainfall, throughfall or streamflow intensity and the infiltration capacity of the soil. Where the input rate exceeds the infiltration capacity of the soil, infiltration excess overland flow will be generated (see Figure 1). In many places, particularly on vegetated surfaces, rainfall rarely exceeds the infiltration capacity of the soil unless the soil becomes completely saturated. Bare soil areas are particularly prone to infiltration excess runoff generation since the energy of the raindrops can rearrange the soil particles at the surface and form a surface crust, effectively sealing the larger pores. A vegetation or litter layer protects the surface and also creates root channels that may act as pathways for infiltrating water. Bare surfaces of dispersive soil materials are particularly prone to crusting and such crusts persist between storms unless broken up by vegetation growth, freeze-thaw action, soil faunal activity, cultivation or erosion. Studies of crusted soils have shown that in some cases infiltration rates after ponding might increase over time more than would be expected as a result of the depth of ponding alone.

During widespread surface ponding, air entrapment and pressure build-up within the soil could have a significant effect on infiltration rates. It has also been suggested that air pressure effects can cause a response in local water tables and that the lifting force due to the escape of air at the surface might initiate the motion of surface soil particles. The containment of air will be increased by the presence of a surface crust of fine material but significant air pressure effects would appear to require ponding over extensive areas of a relatively smooth surface. In the field, surface irregularities (such as vegetation mounds) and the presence of macropores might be expected to reduce the build-up of entrapped air by allowing local pathways for the escape of air to the surface.

In the absence of a surface crust, the underlying soil structure and the macroporosity of the soil is an important control on infiltration rates. Since discharge of a laminar flow in a cylindrical channel varies with the fourth power of the radius, larger pores and cracks may be important in controlling infiltration rates. However, soil cracks and some other macropores, such as earthworm channels and ant burrows, may only extend to limited depths so that their effect on infiltration may be limited by storage capacity and infiltration into the surrounding matrix as well as potential maximum flow rates.

Overland flow may also occur as a saturation excess mechanism. Areas of saturated soil tend to occur first where the antecedent soil moisture deficit was smallest. This will be in valley bottom areas, particularly headwater hollows where there is convergence of flow and a gradual decline in slope towards the stream. Saturation may also occur on areas of thin soils where storage capacity is limited or in low permeability and low slope areas that remain wet during recession periods. The area of saturated soil tends to expand with increased wetting during a storm, and reduce again after rainfall stops at a rate controlled by the supply of water from upslope. This is the dynamic contributing area concept. Any surface runoff on such a saturated area may not be entirely due to rainfall but may also be due to a return flow of subsurface water, and there is some evidence that, even under such conditions of a saturated surface, rainfall may still locally infiltrate into the soil. In this way, surface runoff may be maintained during the period after rainfall has stopped.

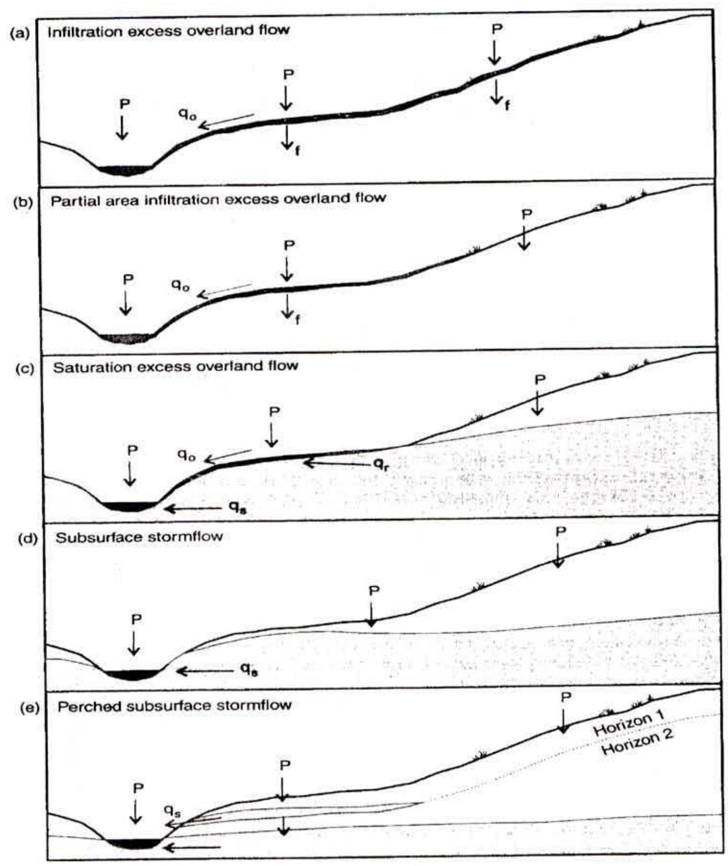

Figure 1: Various hillslope runoff mechanisms.

A similar concept may be invoked in areas where responses are controlled by subsurface flows. When saturation starts to build up at the base of the soil over relatively impermeable bedrock, it will start to flow downslope. The connectivity of saturation in the subsurface will, however, is important initially. It may be necessary to satisfy some initial bedrock depression storage before there is a consistent flow downslope. The dominant flow pathways may be localized, at least initially, related to variations in the form of the bedrock surface. Some catchments, with high infiltration capacities and reasonably deep soils, may have responses dominated by subsurface stormflow. It is worth remembering that a 1m depth of soil, with an average porosity of 0.4 has a storage capacity of 400 mm of water. Thus, if the infiltration capacity of the soil is not exceeded, a large 100 mm rainstorm could, in principle, be totally absorbed by that 1 m soil layer, even if the antecedent storage deficit is only a quarter of the porosity.

It is a common and convenient assumption that the bedrock underlying small upland catchments is impermeable. This is not always the case, even in rocks that have little or no primary permeability in the bulk matrix. The presence of secondary permeability in the form of joints and fractures can provide important flow pathways and storage that may be effective in maintaining stream baseflows over longer periods of time.

It is possible that connected fracture systems that are full of water could act as pipe systems, transmitting the effects of recharge very rapidly. If water is added to one end of a pipe full of water, there will be an almost instantaneous displacement of water out of the other end, whatever the length of the pipe. The reason is that the transmission of the pressure effect of adding the water is very much faster than the actual flow velocity of the water. Such displacement effects explain rapid subsurface responses to storm rainfalls.

The perceptual model briefly outlined above represents a wide spectrum of possible hydrological responses that may occur in different environments or even in different parts of the same catchment at different times. Traditionally, it has been usual to differentiate between different conceptualizations of catchment response based on the dominance of one set of processes over another, for example, the Hortonian model in which runoff is generated by an infiltration excess mechanism all over the hillslopes [Figure 1(a)]. This model is named after Robert E. Horton (1875-1945), the famous American hydrologist. Many catchments are forested with soils that are deeply weathered and have generally high infiltration capacities. Surface runoff is restricted mainly to the channels, so here the storm runoff production must be controlled by subsurface responses [Figure 1(d)].

Betson (1964) suggested that it would be more usual that only part of a catchment would produce runoff in any particular storm and that since infiltration capacities tend to decrease with increasing soil moisture and the downslope flow of water on hillslopes tends to result in wetter soils at the base of hillslopes, then the area of surface runoff would tend to start close to the channel and expand upslope. This partial area model [Figure 1(b)] allowed for a generalization of the Horton conceptualization. It is now realized that the variation in overland flow velocities and the heterogeneities of soil characteristics and infiltration rates are important in controlling partial area responses. If runoff generated on one part of a slope flows onto an area of higher infiltration capacity further downslope it will infiltrate (the run-on process). If the high intensity rainfall producing the overland flow is of short duration, then it is also possible that the water will infiltrate before it reaches the nearest stream channel.

Before discussing about the models of rainfall runoff process, it is helpful to understand the various features of a streamflow hydrograph.

HYDROGRAPH ANALYSIS

A streamflow hydrograph is a graph of the time distribution of water discharge at a location. The graph is plotted with discharge on the ordinate and time on the abscissa. A hydrograph for a given storm reflects the influence of all the physical characteristics of the drainage basin and, to some extent, also reflects the characteristics of the storm causing the hydrograph. A hydrograph can be considered a thumbprint of the drainage basin. The actual shape of a hydrograph is determined by the rate at which water is transmitted from the various parts of the drainage basin to the gauge. Most of this water is carried by the channels, but some water flows overland directly to the gauge.

No two drainage basins produce identical hydrographs for the same storm. Similarly, no two storms produce identical hydrographs from the same basin.

Components of Streamflow

The three main components of runoff are: (a) direct runoff and (b) baseflow. The direct runoff is divided into surface runoff and quick interflow, whereas the baseflow is divided into delayed interflow and groundwater runoff. The division into quick and delayed interflows is essentially arbitrary.

Total runoff corresponds to a given storm event and its volume is determined by including in the streamflow hydrograph all runoff between the baseflow discharge occurring prior to the storm up to the same baseflow discharge after the storm.

Surface Runoff

Surface runoff or overland flow is that water which travels over the ground surface to a drainage channel. Most surface runoff flows to first-order channels because they collectively drain the greatest area of the drainage basin. Surface runoff also includes that precipitation that falls directly on water flowing in the channel. Sheet flow usually occurs from an impervious surface such as a paved parking lot, but can only occur on a natural drainage basin when rainfall intensity uniformly exceeds the infiltration capacity. This condition does not frequently happen. Variations in the distribution of soil type and of rainfall over a drainage basin usually result in limited sheet flow. Surface runoff is believed to be the principal contributor to the peak discharge from a storm event. Because this water runs off over the surface to the channel, it is the first to reach the channel and, hence, forms the rising limb and peak of the hydrograph.

Interflow

Interflow, also called subsurface storm flow, is that surface water that infiltrates the surface layer and moves laterally beneath the surface to a channel. Interflow can occur on forest floors, where the leaves, needles, and other debris cover the ground. Interflow might occur in shallow soils filled and loosened by tree roots, rock debris covering the ground surface, or surface soils loosened by any cause. During interflow, the movement of water is subject to greater flow resistance than surface runoff. As a result, interflow does not move as rapidly as surface runoff. Accordingly, interflow does not add to the peak discharge, but reaches the outlet after the peak discharge has passed.

Direct Runoff

Direct runoff is usually considered to be the sum of surface runoff and interflow. Direct runoff is frequently equated with surface runoff. These two flow components move more rapidly than groundwater flow and for this reason are often lumped together for hydrologic purposes. Such lumping is reasonable for certain purposes because it is logical to believe that some interflow near the outlet will arrive at that point before surface runoff from farther up the basin.

Baseflow

Baseflow, or groundwater flow, is the flow component contributed to the channel by groundwater. Groundwater occurs from surface-water infiltration to the water table and then moving laterally to the channel through the aquifer. Such water moves much more slowly than direct runoff and, for this reason, does not contribute to the peak discharge for a given storm. Flow in a perennial stream prior to a storm is from baseflow. During a storm event, the baseflow is augmented by infiltration. Drainage basins with highly permeable, thick soils usually have a high groundwater-flow component and relatively small direct-flow component, whereas basins with heavy-clay, low-infiltration soil have a small or zero groundwater component and a high direct-runoff component. A portion of the groundwater-flow component occurs from water infiltrating the banks of the channel during high-water flows.

Delineation of Runoff Components

A streamflow hydrograph is shown in Figure 2, with points illustrating the three flow components. By common definition, point A marks the beginning of surface runoff, which is believed to end at the change in slope shown as B; point B is considered to be the beginning of interflow, which ends at point C; point C marks the beginning of groundwater flow, which continues beyond the end of the hydrograph. Of course, this division of hydrograph flow components is subjective and has no quantitative basis. There is no way to verify the source of water during most of the hydrograph and certainly the separation boundaries cannot be verified.

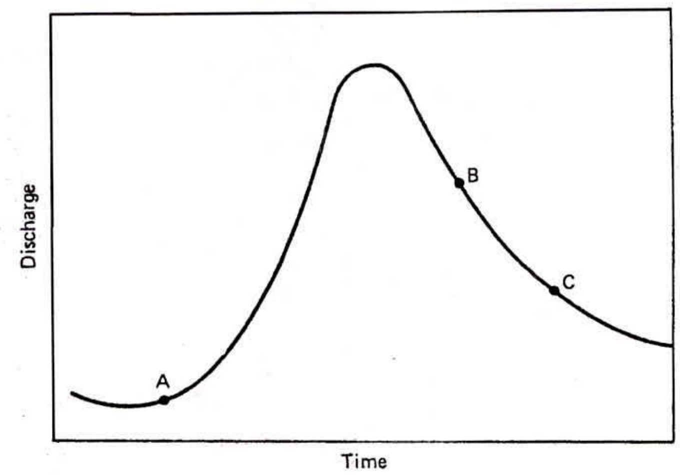

Figure 2: A typical streamflow hydrograph.

Several factors affect a streamflow hydrograph during a runoff event on a drainage basin. These factors are (a) drainage characteristics, (b) rainfall characteristics, and (c) soil and land use.

Elements of the Hydrograph

Typical elements of a hydrograph are shown in Figure 3.

Rising Limb

As surface runoff reaches the gauge, the water begins to rise in the channel. With continuing elapse of time, more and more surface runoff reaches the gauge and the water in the channel continues to rise until it reaches a maximum discharge and after this stage, water begins to recede. The rising portion of the hydrograph is called the rising limb.

Crest

The time interval of the greatest discharge at the peak of the hydrograph is called the crest. The crest might be of a short time interval represented by a sharp peak or of a fairly long interval represented by a flat peak. The crest is not necessarily composed of equal discharges, but rather represents a subjective zone of nearly equal highest discharges. The greatest discharge within the crest is the peak discharge, which is of primary interest in hydrologic design.

Recession Limb

The portion of the hydrograph after the peak is known as the receding limb, falling limb, or recession curve. The receding limb represents decreasing discharge as water is withdrawn from the drainage- basin storage after rainfall ceases. The slope of the receding limb indicates the rate at which water is drained from the basin. The lower part of the recession, which has a much lower slope, is believed to represent groundwater contribution because the water is withdrawn much more slowly than the other components.

Streamflow recession can be expressed as

$$Q_t = Q_0 K_r^t \tag{1}$$

where Q_0 is the initial discharge at any time, Q_t is the discharge at time interval t later, and K_r is the recession or depletion constant r dependent upon the units of time and is less than unity.

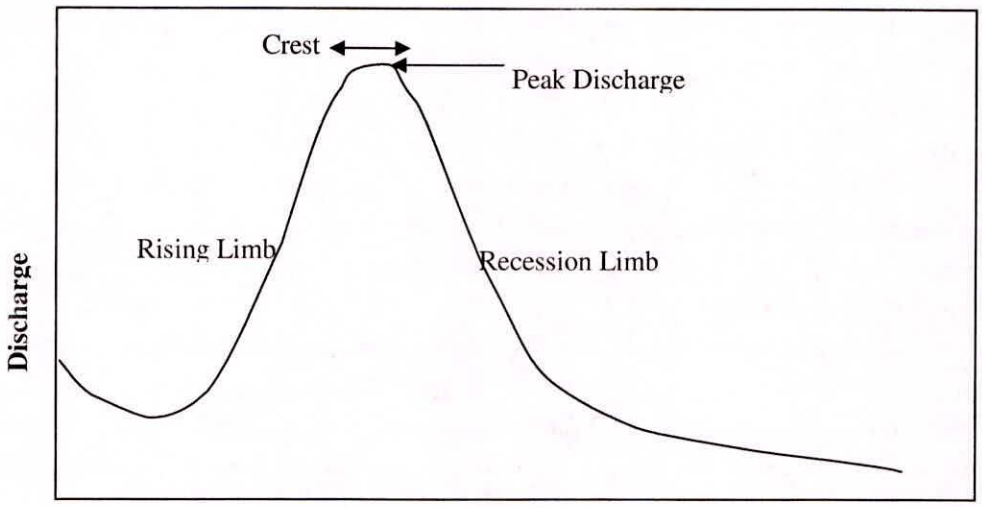

Figure 3: The elements of a hydrograph.

Hydrograph Time Characteristics

The shape, and therefore the characteristics, of a hydrograph can be measured in terms of time.

Time to Peak

The time to peak is the time elapsed from the beginning of the rising limb to the peak discharge. It depends upon the drainage-basin characteristics such as travel distance, drainage density, channel slope, channel roughness, and soil infiltration characteristics. It is altered somewhat by the distribution of rainfall over the basin. For a given amount of runoff, a longer time to peak has a lower peak discharge than a shorter time to peak.

Time of Concentration

The time of concentration is the time required for a drop of water falling on the most remote part of the drainage basin to reach the basin outlet. It includes the time required for all portions of the drainage basin to contribute runoff to the hydrograph and this time then represents the maximum discharge that can occur from a given storm intensity over the drainage basin. By assuming a uniform rainfall over the entire drainage basin, the discharge increases as water from progressively farther distances arrives at the outlet. The hydrograph continues to rise as time elapses and rainfall continues until drainage from the most remote point on the basin arrives at the gauge. At this time, the discharge becomes a constant because all areas within the basin are contributing to the discharge. If the rainfall continues at the same uniform rate, the hydrograph peak would become flat at its maximum discharge and would continue so until rainfall intensity changes. It is important to distinguish between the time to peak and the time of concentration. In practice, the hydrograph peak is sharply defined and the storm duration is less than the time of concentration.

The time of concentration can be determined by many formulas. One of the most commonly used formulas is by Kirpich (1940):

$$t_c = 0.0078\ (L/S^{0.5})^{0.77} \qquad (2)$$

where t_c is the time of concentration in minutes, L is the length of travel in feet from the most remote point on the drainage basin along the drainage channel to the basin outlet, and S is the slope in feet per foot determined by the difference in elevation of the most remote point and that of the outlet divided by L. The equation assumes uniform rainfall over the drainage basin.

RAINFALL-RUNOFF MODELS

A model is a simplified representation of a complex system. It aids in making decisions, particularly where data or information are scarce or there are large-number of options to choose from. Hydrological models represent the physical/ chemical/biological characteristics of the catchment and simulate the natural hydrological processes. Hydrological models are essentially mathematical models where the physical processes of hydrologic cycle are described by a set of mathematical equations, logical statements, boundary conditions and initial conditions, expressing relationships between inputs, variables and parameters. Hydrological models may be broadly classified in two groups:

(i) Deterministic Hydrological Models,
(ii) Stochastic Hydrological Models.

A deterministic hydrological model is one in which the processes are modelled based on definite physical laws and no uncertainties in prediction are admitted. Deterministic models permit only one outcome from a simulation with one set of inputs and parameter values. It has no component with stochastic behaviour, i.e. the variables are free from random variation and have no distribution in probability. Deterministic models can be further classified according to whether the model gives a spatially lumped or distributed description of the catchment area, and whether the description of the hydrological processes is empirical, conceptual or fully physically based.

Stochastic models allow for some randomness or uncertainty in the possible outcomes due to uncertainty in input variables, boundary conditions or model parameters. The vast majority of models used in rainfall-runoff modelling are used in a deterministic way, although again the distinction is not clear-cut since there are examples of models which add a stochastic error model to the deterministic predictions of the hydrological model and there are models that use a probability distribution function of state variables but make predictions in a deterministic way. A working rule is that if the model output variables are associated with some variance or other measure of predictive dispersion the model can be considered stochastic; if the output values are single valued at any time step the model can be considered deterministic, regardless of the nature of the underlying calculations.

Empirical or black box models contain no physically based transfer function to relate input to output. In other words no consideration of the physical processes is involved in such models. These models are basically input-output based models. Within the range of calibration, such models may be highly successful. However, in extrapolating beyond the range of calibration, the physical link is lost and the prediction relies on mathematical technique alone.

Lumped conceptual models occupy an intermediate position between the fully distributed physically based approach and empirical black box analysis. Lumped models treat the catchment as a single unit, with state variables that represent averages over the catchment area, such as average storage in the saturated zone. Such models are formulated on the basis of a relatively small number of components, each of which is a simplified representation of the process element in the system being modelled. Parameters of such type of models are calibrated using trial and error method or automatic optimisation technique or combination of both.

Fully distributed physically based models are based on our understanding of the physics of the hydrological processes which control catchment response and use physically based equation to describe these processes. From their physical basis such models can simulate the complete runoff regime, providing multiple outputs (e.g. river discharge, phreatic surface level and evaporation loss) while black box models can offer only one output. Unlike lumped conceptual models, physically based distributed models do not consider the transfer of water in a catchment to take place between a few defined storages. Instead the transfers of mass, momentum and energy are calculated directly from the governing partial differential equations.

There is a general correspondence between lumped models and the 'explicit soil moisture accounting' (ESMA) models and between distributed models and 'physically based' or process-based models. Even this correspondence is not exact, however, since some distributed models use ESMA components to represent different sub-catchments or parts of the landscape as hydrological response units, while even the most distributed models currently available must use average variables and parameters at grid or element scales greater than the scale of variation of the processes. They are consequently, in a sense, lumped conceptual models at the element scale. There is also a range of models that do not make calculations for every point in the catchment but for a distribution function of characteristics.

A question can arise at this stage as to how choose a particular model structure for a particular application? The following procedure is suggested based, in essence, on considerations of the function of possible modelling structures:

1. Prepare a list of the models under consideration. This list may have two parts: those models that are readily available, and those that might be considered for a project if the investment of time (and money!) appeared to be worthwhile.
2. Prepare a list of the variables predicted by each model and those required. Decide whether the model under consideration will produce the outputs needed to meet the aims of a particular project. If you are interested in the rise in the water table in valley bottoms due to deforestation, for example, a model predicting the lumped response of the catchment may not fulfill the needs of the project. If, however, you are only interested in predicting the discharge response of a catchment for real-time flood forecasting, then it may not be necessary to choose a distributed modelling strategy.
3. Prepare a list of the assumptions made by the model (see the guides in the chapters that follow). Are the assumptions likely to be limiting in terms of what you know about the response of the catchment you are interested in? Unfortunately the answer is likely to be yes for all models, so this assessment will generally be a relative one, or at best a screen to reject those models that are obviously based on incorrect representations of the catchment processes.

4. Make a list of the inputs required by the model, for specification of the flow domain, for the specification of the boundary and initial conditions and for the specification of the parameter values. Decide whether all the information required can be provided within the time and cost constraints of a project.
5. Determine whether you have any models left on your list. If not, review the three previous steps, relaxing the criteria used. If predictions are really required for an application, one model at least will need to be retained at this stage!

RAINFALL RUNOFF MODELLING

A brief description of important model types is given here.

The Rational Method

Rainfall-runoff modelling has a long history and the first hydrologists attempting to predict the flows that could be expected from a rainfall event were had insight into hydrological processes, even if their methods were limited by the data and computational techniques available to them. We can go back nearly 150 years to the first widely used rainfall-runoff model, that of the Irish engineer Thomas James Mulvaney (1822-1892) and published in 1851. The model was a simple equation that manages to illustrate most of the problems that are associated with hydrological modeling. The equation was:

$$Q_p = CAR_m \tag{3}$$

The Mulvaney equation (also known as the rational formula) does not attempt to predict the whole hydrograph but only the hydrograph peak Q_p. This is often all an engineering hydrologist might need to design a bridge or culvert capable of carrying the estimated peak discharge. The input variables are the catchment area, A, a maximum catchment average rainfall intensity, R_m, and an empirical coefficient or parameter, C. Thus, this model reflects the way in which discharges are expected to increase with area and rainfall intensity in a rational way. In fact, variations on equation (3) were published by a variety of authors based on different empirical data sets and are still in use today.

The scaling parameter C reflects the fact that not all the rainfall becomes discharge, but here the method is not quite so rational since it makes no attempt to separate the different effects of runoff production and runoff routing that will control the relationship between the volume of rainfall falling on the catchment in a storm, effectively AR_m, and the discharge at the hydrograph peak. In addition, the coefficient C is required to take account of the nonlinear relationship between antecedent conditions and the profile of storm rainfall and the resulting runoff production. Thus C is not a constant parameter, but will vary from storm to storm on the same catchment, and from catchment to catchment for similar storms. The easiest way to get a value for C is to back-calculate it from observations of rainfall and peak discharge (the very simplest form of model calibration). Predicting the correct value for a different set of conditions, perhaps more extreme than those that have occurred before or for a catchment that has no observations, is a much more difficult task.

Similar difficulties persist to the present day, even in the most sophisticated computer models. It is still difficult to take proper account of the nonlinearities of the runoff production process, particularly in situations where data are very limited. It is still easiest to obtain effective parameter values by back-calculation or calibration where observations are available; it remains much more difficult to predict the effective values for a more extreme storm or ungauged catchment. There are still problems of separating out the effects of runoff production and routing in model parametrizations (and in fact this should be expected because of the real physical interactions in the catchment).

Black box and stochastic models

Black box models attempt to develop empirically identified statistical relationships between rainfall and runoff, without attempting to define and understand the physical processes invoked in the transformation. Generally, these methods require fitting of a mean line through the scatter of plot of runoff against rainfall and use of this mean line or the curve drawn through the points to predict the runoff associated with a particular rainfall. While these models are generally simple to use, they may not be appropriate models of basins which are highly regulated and still under development.

Another class of black box type rainfall-runoff models are the non-parametric Unit Hydrograph based models. These models seek a linear, time invariant and casual relationship between rainfall excess and direct runoff. A variation of this approach is the Linear Perturbation Model (Nash and Barsi, 1983). This class of models establishes a non-parametric linear relationship between total rainfall and total runoff and occupies a special place amongst the group of models called Total Response Linear Models.

Apart from these three model classes, time series methods have also been widely used to design stochastic models of river flow. An essential requirement for the success of this approach is that the processes being modeled must be stationary. This is because these models attempt to preserve the serial correlations in the series being modeled. Where non-stationarity in river flow is principally due to non-stationarity in climatic influences, e.g., rainfall, one could, at least theoretically, use a Transfer Function Noise, TFN, type stochastic model of river flows, in which rainfall is used as an exogenous predictor. Intervention models have also been proposed to incorporate and understand the impact of human influences on river flow. While it may be argued, this class of stochastic models has physical justification, the understanding of water resources utilization within the river basin and the associated elegance of modeling, as permitted by a formal and an explicit water balance approach, will be missing in this approach.

In the context of rainfall-runoff modeling, stochastic models lack the intuitive appeal because these models do not imply the cause and effect relationship between the input and output variables that exists in a typical hydrologic system and have been, with a few exceptions, generally avoided for rainfall-runoff modeling (Todini, 1988). Research in the area of Artificial Intelligence has also resulted in application of techniques based on Artificial Neural Networks to the problem of modeling rainfall to runoff transformation processes and may be classified as a black box model.

Regression Models

Assuming that a general linear relationship with a memory length m exists between rainfall P and runoff Q, runoff may be expressed as:

$$Q_i = a_1 P_i + a_2 P_{i-1} + ...+ a_m P_{i-m+1} + u_i \qquad ...(4)$$

where a_i = regression coefficients, and u_i = the error term. Vector of a values, which are unknown, is estimated by method of least-squares by minimizing the sum of error squares. Having obtained the regression coefficients a, the Q values can be obtained using the above equation.

<u>Monthly rainfall-runoff relationship for gauged catchments</u>

In India more than 80% of the annual rainfall is received in monsoon season (normally from June to October). The rainfall-runoff relationships for monsoon months may be developed using linear rainfall-runoff model. However, during non-monsoon months (Nov-May) most of the runoff in the stream is due to contribution of the ground water reservoir (base flow). The contribution of the rainfall is almost negligible except that a few thunder storms may contribute to streamflow. For partially snow fed basins, snow melt runoff constitutes a significant part of the streamflow. While developing the monthly rainfall runoff relationships, it is necessary to identify the monsoon months for the study area as well as type of the basin i.e. snow fed or rain fed. If the basin is partially snow fed and partially rain fed, monthly snow water equivalents are needed in addition to monthly rainfall data. The form of monthly rainfall - runoff relationships are given below for different conditions.

(a) Monsoon Months

The expression for runoff from large size catchment, in terms of depth of water, can be given in the following form:

$$Q_m = a_1 (P_m - I_{am}) + a_2 (P_{m-1} - I_{am-1}) \qquad .(5)$$

Here, the coefficients a_1 and a_2 are the regression coefficients, and I_{am} represent initial abstraction for month m. Eq. (9) may be expanded to

$$Q_m = a_1 P_m + a_2 P_{m-1} - b_1 I_{am} - b_2 I_{am-1} \qquad .(6)$$

Let $b = -(b_1 I_{am} + b_2 I_{am-1})$, then

$$R_m = b + a_1 P_m + a_2 P_{m-1} \qquad ..(7)$$

The threshold values I_{am} and I_{am-1} cannot be determined exactly. They can only be determined if their relative values are known.

The relationships for non-monsoon months can also be developed based on non-monsoon flows and annual flows. Non-monsoon flow (Q_{NON}) is usually taken as total of runoff for non-monsoon months within a water year. Total runoff for twelve months of a year represent annual flow (Q_{AN}). Two relationships may be obtained in the following steps:

(i) Develop the following relationship between Q_{NON} and Q_{AN}:

$$Q_{NON} = K\, Q_{AN} \qquad (8)$$

(ii) Distribute non-monsoon flows, Q_{NON}, in each of seven months using the following form of relationships:

$$Q_m = K_i * Q_{NON} \qquad .(9)$$

The value of K_i for each of the non-monsoon months may be evaluated as a ratio of average monthly flow for the concerned month to average non-monsoon flow for particular site.

Unit Hydrograph

The unit hydrograph (UH) theory proposed by Sherman (1932) is primarily based on the principle of linearity and time and space invariance. A UH is a hydrograph of surface runoff resulting at a given location on a stream from a unit rainfall excess amount occurring in unit time uniformly over the catchment area up to that location. The rainfall excess excludes losses (abstractions) from total rainfall and unit rainfall excess normally equals 1 mm. The selection of unit time depends on the duration of storm and size of the catchment area. For small catchments, periods of 1 or 2 hours can be assumed and for larger catchments, 3, 4, 6, or even 12 hours can be adopted. Thus

- A UH is a flow hydrograph;
- A UH is a hydrograph of direct surface runoff (DSRO), not total runoff;
- The hydrograph of surface runoff results from the rainfall excess;
- The rainfall excess represents total rainfall minus losses (abstractions);
- During the unit time period, the rainfall excess is assumed to occur uniformly over the

catchment;

- Typical unit times used in UH analyses are 1, 2, 3, 6, 8, and 12 hours. Beyond this, time period is generally taken as an integer multiple of 24 hours.

A unit hydrograph can be interpreted as a multiplier that converts rainfall excess to direct surface runoff. The direct surface runoff (DSRO) is the streamflow hydrograph excluding baseflow contribution. Since, a unit hydrograph depicts the time distribution of flows, its multiplying effect varies with time. In real-world application, the unit hydrograph is applied to each block of rainfall excess and the resulting hydrographs from each block are added for computing direct surface runoff hydrographs, to which baseflows are further added to obtain total hydrographs.

Factors Affecting UH Shape

The factors affecting the shape of the unit hydrograph are the rainfall distribution over the catchment and physiography of the catchment, viz., shape, slope, vegetation, soil type, etc. Variations in areal pattern of rainfall, rainfall duration, and time intensity pattern greatly affects the shape of the hydrograph. A hydrograph resulting from rainfall concentrated in the lower part of a basin will exhibit a rapid rise, sharp peak, and rapid recession. On the other hand, rainfall concentrated in the upper part of the same basin will yield a slow rising and receding hydrograph having broad peak. Thus, UHs developed from rainfall of different areal distributions will exhibit differing shapes. Given the amount of runoff, the time base of the unit hydrograph increases and peak lowers as the duration of rainfall increases.

Natural physical characteristics of a watershed are affected by man's influence, for example, follow-up of watershed management practices significantly change the land cover, and consequently, the shape of the derived UH also changes. Steep catchment slopes produce runoff peak earlier than flatter slopes. Consequently, UHs of steep catchments exhibit peaks occurring earlier than those of flatter slopes. Urbanization of a catchment causes drastic changes in the shape of hydrograph and, in turn, the UH. Urbanisation reduces the natural storage of the basin as well as the average loss rate. As a result, the derived UH exhibits higher peak and shorter time of concentration. Seasonal and long-term changes in vegetation or other causes, such as fire, also changes the physical characteristics of the watershed. It resorts to developing a regional relationship between UH parameters and existing basin characteristics, for deriving the unit hydrograph in the changed environment.

DAILY RAINFALL-RUNOFF RELATIONSHIP

Nash and Barsi (1983) developed a model which relates daily rainfall with daily runoff. The model, originally developed for daily flow forecasting on larger catchments exhibiting seasonality, may also be applied to estimate daily flow corresponding to daily rainfall values. It was assumed that for the year in which the rainfall on each day is exactly the seasonal mean for that day, i_d, the corresponding discharges would also agree with their seasonal means, q_d. Hence

$$i_d \rightarrow q_d \tag{10}$$

It was assumed that in any particular year, the departures of the rainfall and the discharge from these seasonal means are linearly related:

$$i - i_d \rightarrow q - q_d \qquad .(11)$$

or $$x \rightarrow y \tag{12}$$

where,

$$x = i - i_d\,, \qquad y = q - q_d \qquad .(13)$$

The values of i_d and q_d can be obtained by averaging the rainfall and discharge records for each date d over the years in the period of calibration and are smoothened by Fourier analysis. The seasonal values of i_d and q_d are subtracted from the actual values of i and q on each day to obtain the departure series for x and y. Thus, the input and output series for x and y of length equal to the number of days in the calibration period are obtained. Assuming that a general linear relationship with a memory length m exists between the x and y series, y may be expressed as:

$$y_i = h_1 x_i + h_2 x_{i-1} + ...+ h_m x_{i-m+1} + u_i \tag{14}$$

where h_1 = the vector of regression coefficients which represent the discrete series of pulse response and u_i = the disturbance term.

Vector of h values, which are unknown, is estimated by method of least-squares by minimizing the sum of error squares. Having obtained the regression coefficients h, the y values can be obtained using the following equation:

Finally the seasonal mean q_d is added to the y values to give the estimates for q values. The difference between the observed and computed q values provides a series of residual errors for the calibration period. The series of residual errors may be analyzed to identify the following persistence structure:

$$e_i = b_1 e_{i-l} + b_2 e_{i-l-1} + b_3 e_{i-l-2} + ...+ b_n e_{i-l-n+1} + E_i \qquad ...(15)$$

where, l represents the lag period to be identified from the analysis, b_1, b_2 ..b_n are the regression coefficients to be obtained from least square analysis and E_i is the random component with mean zero and standard deviation 1. The estimated daily flow are updated for the residual errors obtained from eq. (15).

CONCEPTUAL MODELS

These models occupy an intermediate position between the fully physically-based approach and empirical black-box analysis. Such models are formulated on the basis of a relatively small number of components, each of which is a simplified representation of one process element in the system being modeled. Models belonging to this group describe catchments as storages which are connected according to a defined rule. Nash Cascade Model, Stanford Watershed Model (SWM), Sacramento model, and Tank Model are some well known conceptual models. In the Tank model model, the catchment is represented by a series of tanks. Recently, Todini (1996) introduced the ARNO rainfall-runoff model. This is a semi-distributed conceptual model using spatial probability distribution of soil moisture capacity and dynamically varying saturated contributing areas. A detailed treatment of lumped models is given by *Blackie and Eeles* [1985] and Singh (1995).

SCS Curve Number Method

The Soil Conservation Service (SCS) of USA has developed a procedure for estimating runoff from small watersheds. This empirical procedure was developed to provide a rational basis for estimating the effects of land treatment and land use changes upon runoff resulting from storm rainfall. Because of its simplicity, however, it has been widely used by agriculturists, hydrologists and soil conservation engineers.

The SCS Curve Number method is widely used because (i) it is a reliable procedure that has been used for many years in different parts of the world, (ii) it is computationally efficient, (iii) the required inputs are generally available, and (iv) it relates runoff to soil type, land use and management practices.

The volume of runoff depends on both meteorologic and watershed characteristics. The precipitation volume is the single most important meteorological characteristics in estimating the runoff. The soil type, land use and the hydrologic condition of the cover are the watershed factors that will have significant effect on the volume of runoff.

The SCS developed an index, which is called the runoff curve number (CN) to represent the combined hydrologic effect of soil, land use, agricultural land treatment class, hydrologic condition and antecedent soil moisture. The curve number, CN, is directly used in the relationship to estimate the runoff.

The volume of runoff (Q) depends on the volume of precipitation (P) and the volume of storage that is available for retention. The actual retention (F) is the difference between the volumes of precipitation and runoff. The main assumptions of the SCS-CN method are: i) runoff begins after an initial abstraction I_a consisting of interception, surface storage, and infiltration has been satisfied, and ii) The ratio of actual retention of rainfall to the potential maximum retention S is equal to the ratio of direct runoff to rainfall minus initial abstraction. Thus

$$(P - I_a - Q) / S = Q / (P - I_a) \tag{16}$$

or

$$Q = (P - I_a)^2 / [(P - I_a) + S] \tag{17}$$

The initial abstraction is a function of land use, treatment, and condition; interception; infiltration; depression storage; and antecedent soil moisture. An empirical analysis was performed by SCS and it was found that 20% of the potential maximum retention is the initial abstraction before runoff starts, or $I_a = 0.2S$. Therefore,

$$Q = (P - 0.2S)^2 / (P + 0.8S) \tag{18}$$

Recent research suggests that the relation $I_a = 0.2S$ may not be correct under all circumstances. However, it remains in use until more reliable results are available. Note that it implies that the factors affecting I_a could also affect S. The parameter S depends upon the soil, vegetation, land use and antecedent moisture condition (AMC) of a catchment. The SCS expressed S as a function of a parameter termed *curve number* :

$$S = 1000 / CN - 10 \tag{19}$$

where CN is the runoff curve number. It represents a measure of retention of water by a soil-vegetation-land use complex. Its permissible range is 0-100. Since S is a function of the factors that affect I_a one can expect that the CN would also be a function of land use, antecedent soil moisture, and other factors that affect runoff and retention. Note that when S is zero, CN is 100 and this leads to Q = P.

To determine the parameter CN, the soil type is divided in four groups. The AMC which represents the moisture content of the soil at a given time has been classified in three groups : AMC I - dry soil, AMC II - medium conditions, and AMC III - saturated soil.

SCS Soil Group Classification

SCS developed a soil classification system that consists of four groups, which are identified by the letters A, B, C and D. Soil characteristics that are associated with each group are as follows:

Group A: deep sand, deep loess, aggregated silts,

Group B: shallow loess, sandy loam,

Group C: clay loams, shallow sandy loam, soils low in organic content, and soils usually high in clay, Group D: soils that swell significantly when wet, heavy plastic clays, and certain saline soils.

SCS Cover Complex Classification

The SCS cover complex classification consists of three factors: land use, treatment or practice, and hydrologic condition. For estimating the curve numbers, approximately fifteen different land uses have been identified. Agricultural land uses are often subdivided by treatment or practices, such as contoured or straight row; this separation reflects the different hydrologic runoff potential that is associated with variation in land treatment. The hydrologic condition reflects the level of land management; it is separated with three classes: poor, fair and good. Not all of the land uses are separated by treatment or condition.

Antecedent Soil Moisture Condition

Antecedent soil moisture is known to have a significant effect on both the volume and rate of runoff. Recognizing that it is a significant factor, SCS developed three antecedent soil moisture conditions, which were labelled I, II and III. The soil conditions for each are as follows:

1. Soils are dry but not to wilting point, satisfactory cultivation has taken place.
2. Average conditions.
3. Heavy rainfall, or light rainfall and low temperatures have occurred within the last 5 days; saturated soil.

The following table gives seasonal rainfall limits for the three antecedent soil moisture conditions:

AMC	Total 5-day Antecedent Rainfall (cm) Dormant Season	Growing Season
I	Less than 1.27	Less than 3.56
II	1.27 to 2.73	3.56 to 5.33
III	Over 2.73	Over 5.33

In design, the antecedent soil moisture condition is often a policy decision rather than a statement of actual soil condition at the site during development.

PHYSICALLY-BASED DISTRIBUTED MODELS

Now-a-days engineers, scientists and planners involved in water resources development have become more concerned with the effect of land use changes related to agricultural and forestry practices, hazards of pollution and toxic waste disposal and general problem arising from conjunctive uses of water. Conventional rainfall runoff models (empirical as well as lumped conceptual models) are often not able to provide satisfactory solutions to such problems. Attention is, therefore, being focused on the physically based distributed catchment models since these have the potential to overcome many of the deficiencies associated with simpler approaches. On the other hand, such models are complex and considerable resources in human expertise and computing capability are needed for their development and applications. Distributed models make predictions that are distributed in space, with state variables that represent local averages of storage, flow depths or hydraulic potential, by discretizing the catchment into a large number of elements or grid squares and solving the equations for the state variables associated with every element grid square. Parameter values must also be specified for every element in a distributed model.

These models simulate or mimic the hydrological processes governing the relationship between rainfall and runoff. Well known examples of this class of models are the European Hydrological System - Systeme Hydrologique European or SHE model (Abbott et al., 1986) and IHDM (Beven, 1987). The SHE system can model all or any part of the land phase of the hydrological cycle. Based, as process models are, on complex physical theory, a high degree of physical realism is sought to be achieved in this approach. An important advantage of developing and using physics based deterministic models is their perceived value in helping to improve our understanding of the hydrologic system. For this reason these models are appropriate tools for modeling dynamics of hydrologic systems in which major changes in land use, in drainage or river control systems occur (Blackie and Eeles, 1985). Jain et al. (1992) applied to SHE model to a few catchments in India.

Beven (1989) wrote a detailed treatise on the scope of physically based models. The paper provides an excellent insight into the as yet unresolved, problems that bedevil this approach. While stating that the theoretical advantages of physically based models remain unproven, the author in his critical evaluation of the physics based approach to hydrologic modeling highlights the following serious drawbacks in this approach. According to Beven (1989), 'the physics on which the equations are based is the small scale physics of homogeneous systems'. In actual applications, lumping of the small scale physics to the model grid scale is required to be done for a numerical solution to the transport equations. There is, however, no theoretical framework for carrying out this lumping of subgrid processes for spatially heterogeneous grid squares. It is merely assumed that the same small scale physics equations can be applied at the model grid scale with the same parameters. This, according to the author, is making a leap into the realm of the conceptual approach. Beven (1989) asserts that it is easy to demonstrate the conceptual nature of current physically based models and states that '... results of the implicit lumping of subgrid processes inherent in physically based models. We cannot be sure that the equations will be the same at the grid scale, nor whether effective grid scale parameters can be defined. For now, it is sufficient to conclude that the current generation of distributed physically based models are lumped conceptual models'. Preliminary studies on lumping of subgrid processes in hydrology seem to suggest that it is not possible to define a consistent effective parameter value to reproduce the response of a spatially variable pattern of parameter values.

The Soil and Water Assessment Tool (SWAT) Model

SWAT is a spatially distributed, continuous time scale watershed model developed by Dr. Jeff Arnold for the USDA-ARS. It was developed to predict the impact of land management practices on water, sediment and agricultural chemical yields in large complex watersheds with varying soils, land use and management conditions over long periods of time. Weather, soil properties, topography, vegetation and land management practices are the most important inputs for SWAT to model hydrologic and water quality in a watershed (Neitsch, 2002).

SWAT model is comprised of numerous diverse physical processes in the basin to be modeled. Catchment has to be divided into sub-catchments for the purpose of modelling. Sub-catchment division in simulations is very useful in the environment with catchment parts having significantly different characteristics of vegetation or soil that has an impact on hydrologic processes. Division of basic catchment areas within the sub-catchments allows the user to distinguish between relevant catchment areas and analyze them. Input data for each sub catchment are grouped or organized into the following categories: climate, HRUs, storages/lakes, underground, river network and catchment runoff. Regardless of the type of problem being modeled and analyzed by the model, background of the method is the water balance of the catchment area. In order to achieve precise forecast of circulation of the pesticides, sediments or nutrients, hydrologic cycle is simulated by the model which integrates overall water circulation in the catchment area. Hydrologic simulations in the catchment area can be divided into two groups. In the soil phase of the hydrologic cycle, the processes on the surface and in the sub-surface soil occur, as well as the circulation of sediments, nutrients and pesticides through the water flows in all sub-catchments. In the second phase, the circulation of water and sediment through the river network up to the exit profile are observed.

SWAT is a semi-distributed, continuous watershed modelling system, which simulates different hydrologic responses using process- based equations. The model computes the water balance from a range of hydrologic processes such as evapotranspiration, snow accumulation, snowmelt, infiltration and generation of surface and subsurface flow components. Spatial variability within a watershed is represented by dividing the area into multiple sub-watersheds, which are further subdivided into hydrologic response units (HRUs) based on soil, land cover and slope characteristics.

SWAT uses a temperature-index approach to estimate snow accumulation and melt. Snowmelt is calculated as a linear function of the difference between average snowpack maximum temperature and threshold temperature for snowmelt. Snowmelt is included with rainfall in the calculation of infiltration and runoff. SWAT does not include an explicit module to handle snow melt processes in the frozen soil, but includes a provision for adjusting infiltration and estimating runoff when the soil is frozen (Neitsch et al., 2005). Despite this limitation, SWAT is considered to be an appropriate integrated model for addressing a range of issues.

Variable Infiltration Capacity (VIC) Model

The variable infiltration capacity (VIC) model was developed for incorporation in GCMs, aiming to improve the representation of horizontal resolution and subgrid heterogeneity in a simple way Gao et al. 2010).

Figure 4 shows the schematic of the VIC model with a mosaic representation of vegetation coverage and three soil layers. The surface of each grid cell is described by N+1 land cover tiles, where n = 1, 2, ... , N represents N different tiles of vegetation, and n = N+1 represents bare soil. For each vegetation tile, the vegetation characteristics, such as LAI, albedo, minimum stomatal resistance, architectural resistance, roughness length, relative fraction of roots in each soil layer, and displacement length (in the case of LAI) are assigned. Evapotranspiration is calculated according to the Penman-Monteith equation, in which the evapotranspiration is a function of net radiation and vapor pressure deficit. Total actual evapotranspiration is the sum of canopy evaporation and transpiration from each vegetation tile and bare soil evaporation from the bare soil tile, weighted by the coverage fraction for each surface cover class. Associated with each land cover type are a single canopy layer, and multiple soil layers (three layers are used for description in this ATBD). The canopy layer intercepts rainfall according to a Biosphere-atmosphere transfer scheme (BATS) parameterization as a function of LAI. The top two soil layers are designed to represent the dynamic response of soil to the infiltrated rainfall, with diffusion allowed from the middle layer to the upper layer when the middle layer is wetter. The bottom soil layer receives moisture from the middle layer through gravity drainage, which is regulated by a Brooks-Corey relationship for the unsaturated hydraulic conductivity. The bottom soil layer characterizes seasonal soil moisture behavior and it only responses to short-term rainfall when the top soil layers are saturated. The runoff from the bottom soil layer is according to the drainage described by the Arno model. Moisture can also be transported upward from the roots through evapotranspiration. Although vegetation subgrid-scale variability is a critical feature for the VIC model, the soil characteristics (such as soil texture, hydraulic conductivity, etc.) are held constant for each grid cell. In the model, soil moisture distribution, infiltration, drainage between soil layers, surface runoff, and subsurface runoff are all calculated for each land cover tile at each time step. Then for each grid cell, the total heat fluxes (latent heat, sensible heat, and ground heat), effective surface temperature, and the total surface and subsurface runoff are obtained by summing over all the land cover tiles weighted by fractional coverage.

The VIC model can be run in either a water balance mode or a water-and-energy balance mode.

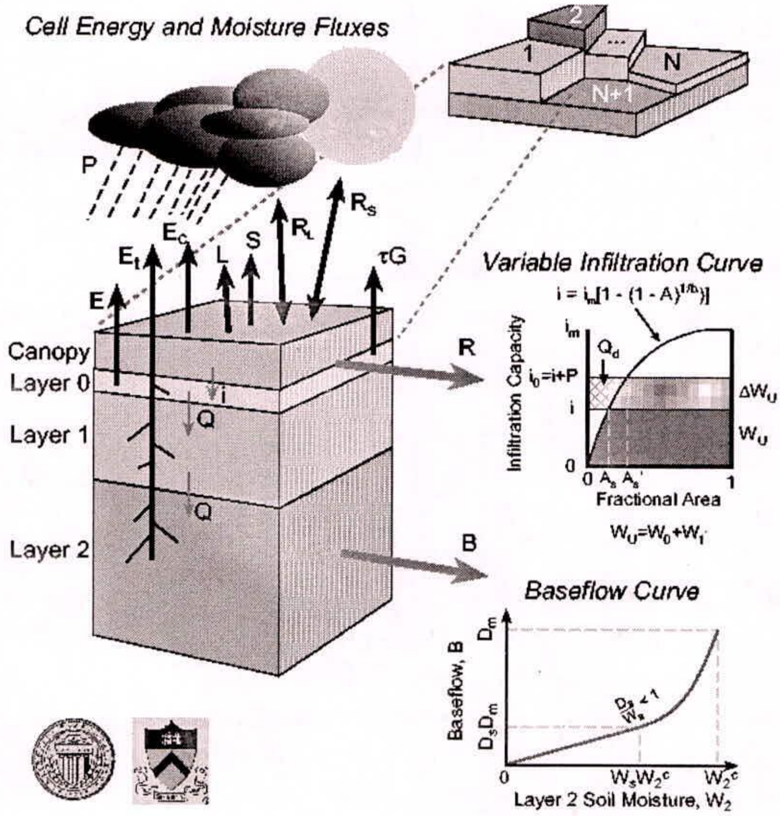

Figure 4 Schematic of the VIC model with mosaic representation of vegetation coverage.

MODEL CALIBRATION AND VALIDATION

Once one or more models have been chosen for consideration in a project, it is necessary to address the problem of parameter calibration. Unfortunately, it is not, in general, possible to estimate the parameters of models by either measurement or prior estimation. Studies that have attempted to do so have generally found that, even using intensive series of measurements of parameter values, the results have not been entirely satisfactory. Prior estimation of feasible ranges of parameters also often results in ranges of predictions that are wide and may still not encompass the measured responses all of the time.

The model performance is typically evaluated from the comparison of simulated and the observed discharge data in terms of mean, standard deviation, maximum daily discharge, and the total discharge using commonly used indices. To evaluate the model performance, three statistical indices are commonly used to evaluate the model performance: Coefficient of determination (R^2), Nash and Sutcliffe efficiency (E), index of agreement (d), and relative error of the stream flow volume (RE). These are computed as follows.

Coefficient of determination (R^2): Willmott (1981) and Leagates and McCabe Jr. (1999):

$$R^2 = \left(\frac{\sum_{i=0}^{n} \left(Y_i^{obs} - Y_{mean}^{obs}\right)\left(Y_i^{sim} - Y_{mean}^{sim}\right)}{\sqrt{\sum_{i=1}^{n} \left(Y_i^{obs} - Y_{mean}^{obs}\right)^2} \sqrt{\sum_{i=1}^{n} \left(Y_i^{sim} - Y_{mean}^{sim}\right)^2}} \right)^2 \qquad \text{....(20)}$$

Nash-Sutcliffe coefficient (E_{NS}) Nash and Sutcliffe (1970) developed this index:

$$E_{NS} = 1 - \left[\frac{\sum_{i=1}^{n} \left(Y_i^{obs} - Y_i^{sim}\right)^2}{\sum_{i=1}^{n} \left(Y_i^{obs} - Y_{mean}^{obs}\right)^2} \right] \qquad \text{....(21)}$$

Index of agreement (d) It was proposed by Willmott (1981) as:

$$d = 1 - \frac{\sum_{i=1}^{n} \left(Y_i^{obs} - Y_i^{sim}\right)^2}{\sum_{i=1}^{n} \left(\left|Y_i^{sim} - Y_{mean}^{obs}\right| + \left|Y_i^{obs} - Y_{mean}^{obs}\right|\right)^2} \qquad \text{.(22)}$$

There are two major reasons for difficulties in calibration. The first is that the scale of the measurement techniques available is generally much less than the scales at which parameter values are required. For example, there may be a hydraulic conductivity parameter in a particular model structure. Techniques for measuring hydraulic conductivities of the soil generally integrate over areas of less than 1 m^2. However, even the most finely distributed models require values that effectively represent the response of an element with an area of 100 m^2 or a much larger area. For saturated flow, there have been some theoretical developments that suggest how such effective values might change with scale, given some underlying knowledge of the fine-scale structure of the conductivity values. In general, however, carrying out the experimental measurements required to use such a theory at the hillslope or catchment scale would be very time-consuming and expensive. Thus it may be necessary to accept that the small-scale values that it is possible to measure and the effective values required at the model element scale are different quantities (a technical word is that they are incommensurate) - even though the hydrologist has traditionally given them the same name. The effective parameter values for a particular model structure will then still need to be calibrated in some way.

Most calibration studies in the past have involved some form of optimization of the parameter values by comparing the results of repeated simulations with whatever observations of the catchment response are available. The parameter values are adjusted between each run of the model, either manually by the modeller or by some computerized optimization algorithm until some 'best fit' parameter set has been found. There have been many studies of different optimization algorithms and measures of goodness of fit or objective functions in hydrological modelling. The essence of the problem is to find the highest peak in the response surface in the parameter space defined by one or more objective functions.

It is highlighted here that the model structure and the observations are not error-free. Thus, the optimum parameter set found for a particular model structure may be sensitive both to small changes in the observations, or to the period of observations considered in the calibration, and possibly to changes in the model structure such as a change in the element discretization for a distributed model. There are a number of important implications that follow from these considerations:

- The parameter values determined by calibration are effectively valid only inside the model structure used in the calibration. It may not be appropriate to use those values in different models or in different catchments.
- The concept of an optimum parameter set may be ill-founded in hydrological modelling. While one optimum parameter set can often be found there will usually be many other parameter sets that are very nearly as good, perhaps from very different part in the parameter space. The idea of equifinality of parameters suggests that, given the limitations of both the model structures and observed data, there may be many representations of a catchment that may be equally valid in terms of their ability to produce acceptable simulations of the available data.

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