1. Mechanisms Producing Event Response

Flow in a channel is the result of excess precipitation on the basin land-surface which is routed down gradient. We will distinguish between

- runoff at the hillslope scale
- runoff routing through the channel network
- overall watershed response which is a combination of both.

We classify the runoff mechanisms to guide our understanding of the mechanisms that lead to runoff or flood response:
-Many of these processes can occur simultaneously in a basin.
-There can be shifts in these mechanisms as a function of time
-Runoff mechanisms may vary in space
-Antecedent conditions determine the temporal variability in runoff processes
-Soils, vegetation, geology, topography and network characteristics play a role in the dominant mechanisms

In general processes involving surface runoff are typically quick (low t_{pk} and intense (high q_{pk} for example channel precipitation and overland flow
Subsurface processes are slow (high t_{pk}) and less intense (low q_{pk}) like subsurface discharge.

Channel Precipitation Rain that falls directly on the stream to become incorporated into channel flow and can be an important contributor to streamflow in some regions.

Overland Flow There are two types of overland flow:

Horton Overland Flow when the sloping surface is saturated from above. Horton (1933, 1945) proposed that rapid runoff results from an excess in the infiltration capacity in a soil due primarily to a low saturated hydraulic conductivity relative to water input. $w > K_{h*}$ and $t_w > t_p$. After ponding, rainfall is converted to surface runoff as $q_{h0}(t) = w(t) - f(t)$, and we can compute f(t) using infiltration equations. Interaction wit the water table is not a pre-requisite.

While originally thought to be the main mechanism it is now recognized to occur for relatively intense rains on relatively fine-grained soils so it is important in semi-arid to arid regions, or regions with relatively impermeable surfaces. This can also occur in on a limited contributing area

Saturation Overland Flow when the sloping surface is saturated from below. As the water table rises to the surface, there is a) direct water input to the saturated area and b) the return flow by the groundwater from upslope. It was established in the seminal work by Dunne and Black (1970). The "break-out" region varies in space and time. Predominates in humid streams that are gaining.

Subsurface Event Flow Flow through the saturated aquifer also occurs during a storm event at a much slower rate. As the water table increases near the stream, so does the head gradient close to the stream, and this produces a flow toward the stream (ground water mounds). This can force "old water" to leave the aquifer and contribute to the hydrography (over and beyond the baseflow).

There can also be flow from perched saturated zones when there is a thin permeable layer over a more impermeable one, then there can be a thin saturated zone when an event occurs and this is referred to as a *sloping slab*. In semi-arid regions, a low permeability calcic horizon could lead to perched saturation during intense thunderstorm events with prior wetness.

Semi-arid areas with very deep water tables are unlikely to have surface-groundwater interactions over broad areas (variable source) expect under conditions of thin perched saturated zones:

(1) Lateral flow can be significant if a perched layer exists

(2) A perched layer creates the conditions for rainfall onto a saturated soil

(3) Macropores in the unsaturated/saturated zone can enhance lateral movement.

Interflow or Throughflow is downslope flow in the unsaturated zone. It is considered to be a minor component in storm event response since the unsaturated zone will have negative matric potentials. Matrix or Darcian flow through the soil matrix is slowbut can occur for long periods of time.

Macropore Flow can play a significant role as runoff conduits on hillslopes. Macopores can form along a) decayed root locations, b) regions of high permeability c) cavities created by animal burrowing. For example, a study in the Los Alamos area by Newman et al. (1999) pointed to interflow as the dominant runoff mechanism.

2. Open Channel flow and Streamflow Routing

Streamflow routing is used to predict an outflow hydrograph at the downstream end of a stream given an inflow hydrograph upstream using the hydraulic characteristics of the stream and the lateral-inflow hydrograph.

2a. *Energy Equation*

The change in head (energy per weight of water) in the x-direction is:

$$\frac{dH}{dx} = S_c - \frac{\partial Y}{\partial x} - \frac{U}{g}\frac{\partial U}{\partial x} - \frac{1}{g}\frac{\partial U}{\partial t} \quad (1)$$

where U is the average velocity, g is acceleration due to gravity, Y is the depth of flow which gives the pressure potential, and $S_c = tan(\beta_c)$ is the tangent of the channel slope. H must decrease downstream due to friction, and all the terms represent forces tending to accelerate the flow. For turbulent flows, the flow rate is proportional to the square root of the energy gradient, and we can use Manning Equation:

$$U = \left(\frac{u_M R^{2/3}}{n}\right)\left(\frac{dH}{dx}\right)^{1/2} \quad (2)$$

Where the *hydraulic radiu* $R = A/P$ where P is the perimeter of the wetted portion

Looking back at equation 1, if the spatial and temporal gradients of depth and velocity are negligible when compared to the slope, Manning's equation reduces to:

$$U = \left(\frac{u_M R^{2/3}}{n}\right) S_c^{1/2} \approx \left(\frac{u_M Y^{2/3}}{n}\right) S_c^{1/2} \quad (3)$$

Because in most channels $R \approx Y$

n characterizes channel conductance and u_M are conversion factors.

2b. *Continuity equation*

The volumetric flow rate through a stream cross section (discharge) $Q[L^3T^{-1} = UA$ for a rectangular channel $Q = UYB$

Mass in $M_{in} = \rho UA + \rho q_L \Delta x$, which is the sum of upstream and lateral inflow. q_L is usually positive because of overland flow but it can be negative in loosing streams or reaches.

Mass out $M_{out} = \rho(U + \partial U/\partial x \Delta x)(A + \partial A/\partial x \Delta x)$
which simplifies to $M_{out} = \rho UA + U\frac{\partial A}{\partial x}\Delta x + A\frac{\partial U}{\partial x}\Delta x$ when neglecting small terms.

Change of Mass in time = $\rho\frac{\partial A}{\partial t}\Delta x$

The full equation then becomes

$$\frac{\partial A}{\partial t} = q_L - U\frac{\partial A}{\partial x} - A\frac{\partial U}{\partial x} = q - \frac{\partial(UA)}{\partial x} = q - \frac{\partial Q}{\partial x} \tag{4}$$

2c. *Flood-Wave Velocity*

$$U_F = \frac{1}{B}\frac{\partial Q}{\partial Y} \tag{5}$$

since

$$Q = \frac{u_M Y^{5/3} S^{1/2} B}{n} \tag{6}$$

then

$$\frac{\partial Q}{\partial Y} = \frac{5}{3}\left(\frac{u_M Y^{2/3} S^{1/2} B}{n}\right) \tag{7}$$

and

$$U_F = \frac{5}{3}U \tag{8}$$

Which means that the flood wave travels with a velocity about 1.67 times that of the water. This equation applies to the flood wave confined to the inside of the channel.

3. Convex Routing Method

The convex model is based on the following relations:

$$QI(t) - QO(t) = \frac{dV(t)}{dt} \tag{9}$$

Where QI is the rate of inflow QO is the rate of outflow and V is the water stored in the reach. We can express the outflow storage relation as:

$$QO(t) = \frac{1}{T^*}V(t) \tag{10}$$

Where $T^* = \frac{X}{U_F}$ is the time it takes for the flood wave to travel through the reach. This is like a linear reservoir. The continuity equation is transformed into discrete form and $\Delta t \leq T^*$, and a routing coefficient $CX = \Delta t$. The outflow at time $i + 1$ is found to be:

$$QO_{i+1} = CX(QI_i) + (1 - CX)QO_i \tag{11}$$

4. Muskinghum Method

We previously presented an example of a conceptual routing model known as the convex routing method. A series of other conceptual and physically-based routing models can be employed to represent a flood wave passage through a channel reach. A channel reach can be considered as a storage element that both delays and affects the total amount of water transmitted. This method models the storage volume of flooding in a river channel by a combination of wedge and prism storages. Assuming that the cross-sectional area of the flood flow is directly proportional to the discharge at the section, the volume of prism storage is equal to KQ where K is a proportionality coefficient which is equal to the time of travel of the flood wave through the channel reach. The volume of wedge storage is $KX(I - Q)$, where X is a weightling factor having the range $0 \leq X \leq 0.5$ and usually is 0.2 in natural systems. The total storage is then:

$$S = KQ + KX(I - Q) = K(XI + (1 - X)Q) \tag{12}$$

For hydrologic routing, the values of K and X are assumed to be specified and constant throughtou the range of flow. The change in storage is:

$$S_{j+1} - S_j = K[(XI_{j+1} + (1 - X)Q_{j+1}) - (XI_j + (1 - X)Q_j)] \tag{13}$$

$$\frac{I_j + I_{j+1}}{2} - \frac{Q_j + Q_{j+1}}{2} = \frac{S_{j+1} - S_j}{\Delta t} \tag{14}$$

We can substitute the storage input/output relation into the continuity equation and obtain the Muskingum Routing Equation to predict the outflow hydrograph (Q_{j+1}) from the inflow hydrograph:

$$Q_{j+1} = C_1 I_{j+1} + C_2 I_j + C_3 Q_j \tag{15}$$

where Co, C1 and C2 are:

$$C_1 = -\left(\frac{Kx - 0.5\Delta t}{K - Kx + 0.5\Delta t}\right) \tag{16}$$

$$C_2 = \left(\frac{Kx + 0.5\Delta t}{K - Kx + 0.5\Delta t}\right) \tag{17}$$

$$C_3 = \left(\frac{K - Kx - 0.5\Delta t}{K - Kx + 0.5\Delta t}\right) \tag{18}$$

(19)

$C_o + C_1 + C_2 = 1$

If observed inflow and outflow hydrographs are available for a river reach, the values of K and X can be determined.