to changing inputs of water through net recharge. In catchments, the groundwater discharge to a stream channel may lag the occurrence of precipitation by days, weeks, or even years. When subsurface water flowing downslope to the stream enters the saturated area near the stream, some of the water is forced to reemerge onto the ground surface because the capacity of the soils and rocks to transmit all of the flowing water downslope is insufficient. This reemerging subsurface water is known as return flow (Figure 10.5b). Like saturation-excess overland flow, return flow can be quite rapid in some catchments with shallow water tables where groundwater mounds may form which dramatically increase local hydraulic gradients toward the stream (Freeze, 1972a, 1972b; Winter, 1983). As was the case with subsurface stormflow, groundwater flow can contribute to either the quickflow or baseflow components of a stream hydrograph, although in most instances the contribution to baseflow is dominant.

10.4.5 “Old” and “new” water

As we have noted (Section 10.3), for streams in forested upland catchments in temperate climates, the bulk of quickflow (even at the peak of a hydrograph) has the chemical composition of water that was resident in the subsurface prior to the storm event. That is, the bulk of quickflow is “old” water, indicating that water contained in soils and rocks in a catchment must be able to reach the stream quickly under storm conditions. This observation may seem at first glance to be at odds with the physical processes described above in which “old” groundwater is viewed as contributing primarily to the (slow) baseflow portion of the hydrograph. There are a number of physical processes that can be called upon to explain the observation that a large fraction of quickflow is “old” water. Consideration of the physics of soil water suggests that processes that cause “new” water delivered to the surface of a catchment to displace “old” water and force it into the stream are common. For example, even if only a small amount of infiltrated water reaches the capillary fringe quickly (e.g., through flow in macropores, especially in the riparian, or near-stream, area), it can change the negative capillary-pressure head in the capillary fringe to a positive pressure head. Such rapid increases in positive pressure head can force “old” groundwater into the stream rapidly. Another mechanism that contributes to the delivery of “old” water to a stream under storm conditions is the entrainment of “old” soil water by overland flow. Overland flow in riparian areas occurs in patches, with surface water flowing into the soil, mixing with shallow soil water, and then reemerging on the surface. The resulting overland flow contains a portion of the “old” soil water with which the “new” water has mixed. The conceptual explanations offered above appear to be consistent with observations, but by themselves do not provide quantitative estimates for the various flows. Quantification requires consideration of inflow, outflow, and storage of water within a catchment.

10.5 Contributing Area and Topographic Controls on Saturation

As we saw in Chapter 7, the topography of the landscape exerts an enormous influence on the movement of water in the subsurface. Topography likewise should control the development of areas of surface saturation and runoff. If we could break a catchment up into blocks (“reservoirs”), we might be able to use the conservation of mass equation
to determine the degree of saturation and potential for runoff generation for each one. Each block would differ in its position along the hillslope and in the slope of the land surface (and probably the water table) through the block. Consideration of inflows, outflows, and runoff potential for all of the blocks in a catchment could provide the starting point for routing water through the catchment.

10.5.1 Contributing area

The degree of saturation of each catchment "block" depends on its water balance. If the inflow to the upslope face of the block from higher portions of the catchment is greater than the outflow from the downslope face of the block, the water table within the block will increase. The inflow rate for a catchment block depends on the contributing area, the area of catchment upslope from a given block that contributes inflow to that block. Contributing area, \( A \), depends on the distance to the divide above the block as well as whether it is convergent, divergent, or planar (Figure 10.6).

To define \( A \), elevation contours are drawn at a specified contour interval (for example, 10 m) for the catchment. Beginning at the base of the catchment, lines are drawn perpendicular to each contour they cross, forming a network of curves similar to the flow nets described in Chapter 6. The lines perpendicular to the contour lines represent flow lines. In areas of the catchment with a uniform slope (planar sections), the flow lines will have a constant spacing (Figure 10.7). In areas where the hillslope is concave, the flow lines tend to converge as one follows them downslope. The surface soil layer between two flow lines on a concave, or convergent, slope is something like a converging channel. As the upstream subsurface flow gets funneled into a smaller area, its depth increases and, with sufficient supply, can saturate the soil. The opposite happens in a divergent section. The increasing distance between flow lines allows the subsurface flow to spread out and thin.

While inflow to a catchment block is proportional to contributing area, local slope, \( \tan \beta \), controls outflow from the block. For example, if the topographic and water table slopes of a block are relatively flat, the hydraulic gradient is small and Darcy's law indicates that water movement will be relatively slow in the absence of changes in hydraulic conductivity. Therefore, we might expect a small outflow and an increase in water storage through time within that block depending on the volume of inflow. The increase in storage is even greater if the block is at the base of a convergent hillslope, such that a great deal of upslope flow into the block occurs. If the water table reaches the surface, the block is completely saturated and any additional water supplied to the block will run off as saturation excess overland flow.

10.5.2 Topographic index

As described above, the important characteristics of a hillslope that influence the likelihood of areas of saturation and runoff developing are the upslope contributing area per unit contour length, \( a = A/c \), and the local slope of the block, \( \tan \beta \). These can be related to each other as discussed in Section 10.5.3, or they can be combined into a single variable that quantitatively captures the effect of topography, such as the topographic index:

\[
TI = \ln(a / \tan \beta).
\]  

(10.5)
ion for each one.

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\[ 10.5 \]
and resolution of the digital elevation data. Accurate identification of the channel network, in particular, depends on using high-resolution elevation data. The current state of the art technique for generating high-resolution topographic data is light detection and ranging (lidar). In fact, the critical need for good topographic data to define channels and channel networks has led to calls for lidar maps to be produced to allow accurate mapping of flood plains (NRC, 2007).

A map of topographic indices for a catchment reveals areas where runoff processes such as saturation-excess overland flow are likely to occur (Figure 10.8). High values of the topographic index indicate areas with large contributing areas and relatively flat slopes, typically at the base of hillslopes and near the stream. These areas also correspond with expected groundwater discharge areas (Chapter 7). Low TI values are found at the tops of hills, where there is relatively little upslope contributing area and slopes are steep. These areas correspond generally with groundwater recharge areas.
of the channel network, the current state of light detection and to define channels to allow accurate runoff processes (e.g.,). High values of and relatively flat areas also correspond to areas found at the top and slopes are steep.

Figure 10.8 Topographic indices for a catchment in Shenandoah National Park. The spatial pattern (a) indicates a likelihood of saturation in the central valley of the catchment. The distribution of values (b) is used in TOPMODEL calculations as described in Section 10.6.
10.5.3 Hillslope stability

The hillslopes of catchments are not static features of the landscape. Over time, hillslopes change slowly due to erosion by runoff and rapidly due to landslides (also known as debris flows). Landsliding on steep hillslopes typically occurs in localized areas where the soils are saturated and the water pressure in the pore spaces is high. Surface runoff most often occurs when soils become saturated so that precipitation can no longer infiltrate (saturation-excess overland flow). If the surface runoff is deep enough and/or the slope steep enough, the flow can dislodge and carry soil particles from the hillslope to the channel, resulting in erosion of the hillslope.

Dietrich et al. (1992) combined simple expressions describing the thresholds of saturation-excess overland flow, landsliding, and hillslope erosion with detailed digital elevation data and careful field observations to predict locations within a catchment where each of these processes dominates. Assuming a constant transmissivity $T$ of the surface soil layer, the saturated subsurface soil discharge across a contour line of length $c$ is given by Darcy's law as

$$Q_{\text{subsurface}} = Tc \tan \beta.$$  \hspace{1cm} (10.6)

The water-table slope is assumed to be equal to the surface slope. The total amount of water reaching the length of contour ($c$) over a specified period of time is $Aq_{\text{total}}$, where $q_{\text{total}} = R$ (the recharge rate, [L T$^{-1}$]) and $A$ is the upslope contributing area. In other words, $q_{\text{total}}$ is the volume of water per unit surface area (or depth) that is moving through the hillslope per unit time. The difference between total runoff past a contour interval ($q_{\text{total}}$) and saturated subsurface discharge ($Q_{\text{subsurface}}$) is saturation-excess overland flow. Thus, overland flow occurs when:

$$Aq_{\text{total}} > Tc \tan \beta$$  \hspace{1cm} (10.7)

or

$$\frac{A}{c} > \frac{T}{q_{\text{total}}} \tan \beta$$  \hspace{1cm} (10.8)

Erosion by overland flow will only occur in the parts of the catchment where the overland flow is deep enough (large specific contributing area $A/c$) or the slope is steep enough (large $\tan \beta$) for the flow to dislodge the soil grains. Dietrich et al. (1992) proposed the following expression for the erosion threshold:

$$\frac{A}{c} > \frac{T}{q_{\text{total}}} \tan \beta + \frac{\alpha}{q_{\text{total}}(\tan \beta)^2},$$  \hspace{1cm} (10.9)

where $\alpha$ [L$^2$ T$^{-1}$] characterizes the resistance of the soil to erosion.
Cohesionless material on a sloping surface becomes unstable, leading to shallow landsliding, when the slope of the surface exceeds a critical value dependent on the soil and water properties and the degree of saturation described by:

\[
\tan \beta > \left( \frac{\rho_s - \sigma \rho}{\rho_s} \right) \tan \phi_f,
\]

(10.10)

where \( \rho \) is water density (i.e., 1000 kg m\(^{-3} \)), \( \rho_s \) is soil density at saturation (of the order of about 2000 kg m\(^{-3} \)), \( \sigma \) is the degree of saturation of the soil mantle, and \( \phi_f \) is the internal angle of friction (a parameter expressing the shear strength due to friction among soil grains). When the soil is saturated (\( \sigma = 1 \)), this reduces to \( \tan \beta > 0.5 \tan \phi_f \) for typical values of soil and water density. When the soil is unsaturated (\( Q_{\text{total}} < Q_{\text{subsurface}} \)), \( \sigma = Q_{\text{total}} / Q_{\text{subsurface}} = Q_{\text{total}} / (T \tan \beta) \).

These expressions for thresholds of saturation overland flow (Equation 10.8), erosion (Equation 10.9), and landsliding (Equation 10.10) all depend on the specific contributing area \( A/c \) and the slope \( \tan \beta \). A plot of the curves defining each threshold in terms of these parameters shows their relationship to each other and the topographic parameters. In Figure 10.9, these threshold curves are plotted for a total runoff \( q_{\text{total}} = 50 \text{ mm day}^{-1} \), assuming \( T = 10^{-4} \text{ m}^2 \text{ s}^{-1} \), \( \phi_f = 35^\circ \), and \( \alpha = 8 \times 10^{-4} \text{ m}^2 \text{ s}^{-1} \), which produces good agreement between predictions and observations of hillslope hydrologic characteristics in a small (1.2 km\(^2 \)) northern California catchment studied by Dietrich et al. (1992).

A diagram such as Figure 10.9 can be used to determine areas of a catchment that are susceptible to erosion and landsliding, which may serve a variety of purposes including guiding land-use decisions. The threshold most susceptible to land-use practices is the erosion threshold. The value of the parameter \( \alpha \) characterizing the resistance of the soil to erosion decreases rapidly with removal of vegetative soil covers and soil disturbance. As the value decreases, the erosion threshold shifts to the left in Figure 10.9, resulting in a larger portion of the catchment that is prone to erosion.

### 10.6 Routing Water through a Catchment Using Catchment Models

The question that we address in this section is how precipitation can be routed through a catchment to calculate streamflow. The most straightforward way to use the theory developed in previous chapters to solve the catchment routing problem is to link together equations for overland flow (e.g., Manning's equation), for flow in the unsaturated zone (e.g., Richards' equation), and for flow in the water-table aquifer using an equation for groundwater flow. The earliest application of such a model was presented by Freeze (1971, 1972a, 1972b), who examined the runoff responses of a hypothetical hillslope to precipitation inputs. More recently with advances in computational power, this approach has been used successfully to simulate flows through catchments in great detail. For example, the Penn State Integrated Hydrologic Model (PIHM, Qu and Duffy, 2007; www.pihm.psu.edu) has been applied at several catchments to study coupled hydrological-biogeochemical processes.
Here we will consider a slightly different and simpler approach using the concepts of catchment “blocks” and the topographic index described in Section 10.5. One framework that uses this approach is known as TOPMODEL, a catchment model that is based on the idea that topography exerts a dominant control on flow routing through upland catchments (Beven and Kirkby, 1979). TOPMODEL uses the equation for conservation of mass (“inflow rate minus outflow rate equals rate of change of storage”) for several “reservoirs” in a catchment—for example, an “interception reservoir” and a “soil reservoir” (Figure 10.10). Rainfall provides the input to the interception reservoir, which is taken to have a capacity of a few mm of water depending on the vegetation type (see discussion of interception in Chapter 2). The outputs from the interception reservoir are evaporation, calculated using an evaporation formula (see Chapter 2), and throughfall, which then forms the input to the soil reservoir. The conservation of mass equation again provides a method for calculating the water balance for the soil reservoir. By linking together the water
The concepts of the TOPMODEL concept provide a method to calculate
balance equations for all of the hypothetical reservoirs in the catchment, a routing com-
putation can be completed.

TOPMODEL performs the bookkeeping for the water balance computations in the
framework of topographically defined elements and uses Darcy’s law to calculate flow
rates through the soil. Consider a segment of a catchment defined by a cut along an ele-
vation contour line at the bottom, and “sides” running perpendicular to contours up to
the catchment divide (Figure 10.11). Recall our assumption that flow is driven by topo-
graphy; hence, this segment is just a portion of a flow net for the catchment. The flow of
subsurface water is conditioned strongly by the local topography. The degree of conver-
gence of “flow lines” (lines perpendicular to the contours) determines how much upslope
area drains to a unit length of contour at any given point. The local slope, the thickness
of the soil, and the hydraulic conductivity of the soil determine the “ability” of the soil
to move water farther down the slope once it has arrived at the given point. Source areas
for surface runoff occur where subsurface water accumulates—points to which large
upslope areas drain (such as convergent hillslopes or “hollows”) and where the capacity
to drain the water downslope is limited (where slopes flatten at the base of hollows).
Figure 10.11 The water balance for a catchment hillslope segment. Throughfall at rate $p$ falls on the segment of area $A$ and thickness $D$. A portion, $R$, of this recharges the subsurface. Subsurface flow from the segment occurs at rate $q_{\text{subsurface}}$. Surface flow, $q_{\text{overland}}$, occurs from saturated areas (saturation-excess overland flow). The local slope at the outflow point, $\beta$, is considered to be equal to the slope of the water table.

Conservation of mass can be applied to the segment depicted in Figure 10.11 to determine the fluxes.

10.6.1 TOPMODEL calculations

Streamflow is the sum of subsurface flow and of overland flow from saturated contributing areas:

$$q_{\text{total}} = q_{\text{subsurface}} + q_{\text{overland}}$$  \hspace{1cm} (10.11)

where $q_{\text{total}}$ is total streamflow. It has dimensions of [L T$^{-1}$] (discharge [L$^3$ T$^{-1}$] divided by area [L$^2$]); all of the flow quantities in TOPMODEL have these units. The surface flow contribution is $q_{\text{overland}}$ and $q_{\text{subsurface}}$ is the subsurface flow contribution (Figure 10.11).

Surface flow is generated when precipitation falls on a saturated area and from return flow, so:

$$q_{\text{overland}} = \frac{A_{\text{sat}}}{A} p + q_{\text{return}}$$  \hspace{1cm} (10.12)

where $A_{\text{sat}}/A$ is the fraction of the hillslope area that is saturated (Figure 10.11), $p$ [L T$^{-1}$] is the throughfall or snowmelt rate, and $q_{\text{return}}$ [L T$^{-1}$] is the return flow.

We calculate $q_{\text{subsurface}}$ [L T$^{-1}$] from total subsurface discharge $Q_{\text{subsurface}} = T \tan \beta$ [L$^3$ T$^{-1}$] (Equation 10.6), where $T$ is the transmissivity of the soil [L$^2$ T$^{-1}$], $c$ [L] is the
contour width (length perpendicular to the flow direction), and \( \tan \beta \) is the slope. The transmissivity is equal to the soil depth multiplied by the soil hydraulic conductivity. Note that the slope of the water table is assumed to be the same as that of the land surface.

We assume that the saturated hydraulic conductivity of the soil decreases with soil depth exponentially, a situation often observed:

\[
K(z) = K_0 e^{-z \rho f}.
\]  

(10.13)

where \( K(z) \) [L T\(^{-1}\)] is the hydraulic conductivity at depth \( z \) (measured positively in the downward direction), \( K_0 \) is the hydraulic conductivity at the surface, and \( f \) [L\(^{-1}\)] is a parameter that governs the rate of decrease of \( K \) with depth. To determine the transmissivity of a saturated zone of a given thickness (from a depth of water table \( z \) to a depth to bedrock \( D \)) Equation 10.13 is integrated to obtain:

\[
T = \frac{K_0}{f} (e^{-z \rho f} - e^{-D \rho f}).
\]  

(10.14)

The term \( e^{D \rho f} \) is generally much smaller than the term \( e^{z \rho f} \), so Equation 10.14 can be simplified:

\[
T = \frac{K_0}{f} e^{-z \rho f}.
\]  

(10.15)

Combining Equations 10.6 and 10.15 gives the following equation for subsurface flow:

\[
Q_{subsurface} = \frac{K_0}{f} e^{-z \rho f} c \tan \beta.
\]  

(10.16)

TOPMODEL does water-balance accounting by keeping track of the “saturation deficit,” the amount of water that one would have to add to the soil at a given point to bring the water table to the surface. Because one has to track saturated areas if saturation-excess overland flow is to be computed, this makes sense. To implement computations in terms of \( s \), the saturation deficit, \( z \) is replaced by \( s/\phi \) where \( s \) [L] is the saturation deficit and \( \phi \) is the porosity of the soil. Substituting for \( z \) in Equation 10.16 gives:

\[
Q_{subsurface} = \frac{K_0}{f} e^{-s/\phi} c \tan \beta.
\]  

(10.17)

To make things “neater,” we introduce some simplifying notation. We can replace \( K/\phi \) with \( T_{\text{max}} \) because this term is the transmissivity when the soil is completely saturated \( (s = 0) \). We can also replace \( f\phi \) with \( 1/m \), a soil parameter inasmuch as \( f \) defines the decrease of \( K \) with depth and \( \phi \) is porosity. Equation 10.17 can then be written:

\[
Q_{subsurface} = T_{\text{max}} e^{-s/m} c \tan \beta.
\]  

(10.18)
We may now proceed with a calculation of the water balance for a hillslope slice (Figure 10.11). This will lead, in conjunction with Equation 10.18, to expressions for $\frac{A_{sat}}{A}$ and $q_{subsurface}$. Equation 10.18 gives the subsurface flow being transmitted downslope at any point. The flow coming into the slice at any time is:

$$Q_s = RA,$$

(10.19)

where $R$ [L T$^{-1}$] is the recharge rate and $A$ is the area of the hillslope slice—the section of hillslope that drains past the section of contour (c) in question. The “great leap” of TOPMODEL is to assume steady-state conditions. Then, $Q_{subsurface} = Q_{R}$, or

$$RA = T_{max} e^{-\frac{1}{m_c \tan \beta}}.$$

(10.20)

This equation can be solved for $s$:

$$s = -m \ln \left( \frac{R}{T_{max}} \right) - m \ln \left( \frac{a}{\tan \beta} \right),$$

(10.21)

where $a = A/c$, the specific contributing area. The second term on the right of Equation 10.21 describes the way in which topography controls the propensity for every point in the catchment to reach saturation (i.e., the propensity of each point to generate saturation-excess overland flow during storms). If $s$ is less than or equal to zero, the soil is saturated. From Equation 10.21 we see that this occurs most easily for points within the catchment where the topographic index $(Tl = \ln(a/\tan \beta))$ is large.

Until this point in the discussion, we have been referring to an individual catchment hillslope, or hillslope segment, defined by a pair of streamlines and extending from the stream to the catchment divide (Figure 10.11). However, we could also consider any point in the catchment and calculate the upslope contributing area and the local slope. In this way, we can compute the distribution of topographic indices for the entire catchment. In practice, the computations are often done for “blocks” delineated on the basis of DEM (Digital Elevation Model) or surveying data. The topographic index, and therefore the contributions of surface and subsurface flow to streamflow, can be calculated for each block. The saturation deficit can also be calculated for each block, using Equation 10.21. Furthermore, these quantities will be identical for two blocks with the same topographic index, as long as $R$ and $T_{max}$ are spatially constant.

To solve for the catchment-average saturation deficit ($\bar{s}$), we can integrate Equation 10.21 over the catchment and divide by the area. Here we assume that $R$ and $T_{max}$ are constant over the catchment:

$$\bar{s} = -m \ln \left( \frac{R}{T_{max}} \right) - m \bar{\lambda},$$

(10.22)

where $\bar{\lambda}$ is the mean $\ln(a/\tan \beta)$ for the catchment. Combining Equations 10.21 and 10.22 gives:
\[ s = \overline{r} + m \left[ \lambda - \ln \left( \frac{a}{\tan \beta} \right) \right]. \quad (10.23) \]

This equation states that the saturation deficit at any point in a catchment is equal to the average saturation deficit for the catchment plus a soil parameter, \( m \), times the difference between the average topographic index and the local topographic index.

Now we have a way to calculate \( A_{\text{sat}}/A \)—compute \( s \) at any point and check to see if it is less than or equal to zero. We can estimate \( m \) from soil characteristics, calculate \( \lambda \) and \( \ln(a/\tan \beta) \) from a topographic map, and can keep track of \( \overline{r} \) by water-balance accounting (\( p \), interception, \( e_t \), subsurface flow, and overland flow). If \( s < 0 \), the soil is completely saturated and any rain on the surface will become overland flow. The rate of flow produced by this mechanism is determined using the throughfall intensity and fractional catchment area that is saturated (e.g., Equation 10.12). Return flow occurs where \( s < 0 \), and the rate of return flow is equal to \( |s| A_{\text{sat}}/A \).

Next we develop an expression for the mean subsurface discharge, \( \overline{q}_{\text{subsurface}} \). Integrating Equation 10.18 over the catchment area and dividing the result by the catchment area yields:

\[ \overline{q}_{\text{subsurface}} = T_{\text{max}} e^{-\lambda} e^{-\overline{r}/m}. \quad (10.24) \]

Thus, in TOPMODEL, subsurface flow is controlled by soil characteristics (\( T_{\text{max}} \) and \( m \)), topography (\( \lambda \)), and the average saturation deficit of the catchment.

This is not all of TOPMODEL, but it is a summary of the main conceptual points. Added to the formulations above are the other standard components of the water budget—evapotranspiration, snowmelt, channel routing (e.g., a simple reservoir routing method). All of the water-balance accounting parts of the model are simple applications of the conservation of mass. A fuller description of TOPMODEL is available in Wolock (1993).

### 10.6.2 Hydrograph simulation using TOPMODEL

TOPMODEL simulates the runoff response of a catchment to precipitation events. By tracking the change in storage within each “block” defined by a value of \( T_l \), the model not only routes water through the catchment, but enables predictions of areas that will become saturated during a storm. The soils at the base of hillslopes (blocks with high \( T_l \)s) tend to become saturated as a storm progresses and saturation-excess overland flow is produced. Streamflow is taken as the sum of subsurface flow (net outflow from the soil reservoir) and the overland flow. Evapotranspiration removes water from the soil store according to a rate calculated with an evapotranspiration formula.

To illustrate the use of TOPMODEL to simulate streamflow from a catchment, we examine the Snake River (Hornberger et al., 1994). The catchment of the Snake River near Montezuma, Colorado, is 12 km² in area and is mountainous, ranging in elevation from about 3350 m to 4120 m. The Continental Divide bounds the catchment on the south and east. Approximately half of the catchment is above the tree line. The data necessary to implement the model were taken from nearby stations. A snowmelt model that takes
Figure 10.12 Observed discharge (1984) in the Snake River, Colorado, compared with TOPMODEL calculations.

Data courtesy of Ken Bencala and Dianne McKnight.

Melt to be proportional to temperature was used in the water-balance calculations within TOPMODEL.

The general shape of the snowmelt hydrograph for 1984 simulated using TOPMODEL is in accord with the measured hydrograph (Figure 10.12). The initial timing of the hydrograph rise is late in the simulated hydrograph; presumably, the temperature index model calculates the initiation of significant snowmelt to be later than when the melt is actually initiated. This likely is due to errors in the simple model that was used for snowmelt. The remainder of the simulation follows the observed hydrograph quite well although the late part of the recession is overpredicted. Overall this simulation is quite reasonable given the limitations in the data (e.g., the need to extrapolate the discharge, precipitation, and temperature data from downstream stations).

10.7 Concluding Remarks

At the beginning of the chapter we suggested that solving the catchment routing problem was important for a number of reasons. The material presented above introduced some of the ideas used in routing flows through a catchment and discussed some of