Brutsaert (2005)

"Hydrology, an Introduction"

11 STREAMFLOW GENERATION: MECHANISMS AND PARAMETERIZATION

Streamflow is one of the main manifestations of the hydrologic cycle in nature. It is normally characterized by a hydrograph, that is the rate of flow in the stream channel as a function of time,

$$Q = Q(t) \tag{11.1}$$

A streamflow hydrograph at any point along a river is the integrated result of all flow processes upstream in the catchment, in response to precipitation, and possibly to snowmelt and other water inputs. Therefore, streamflow is not a local but a basin-scale phenomenon. In the previous chapters some of the more important transport phenomena and their mathematical formulations, that are amenable to analysis, have been considered. Most of these mechanisms are fairly well understood individually. However, at present there is still no unifying theory available that provides a coherent and satisfactory explanation for the integration of these different local mechanisms into the streamflow generation process. The main reason for this uncertainty is undoubtedly the large variation in drainage basins; each drainage basin behaves in many respects almost as if it were a law unto itself, and this has made it difficult to derive general relationships that are broadly applicable. But even for any given basin, it is often difficult to identify and quantify the different mechanisms that produce the observed $Q(t)$; the decomposition of an integral into its constituent parts, that is its inversion to obtain the integrands, like "unscrambling an omelet," is not a simple matter.

11.1 RIPARIAN AREAS AND HEADWATER BASINS

The transformation of precipitation, after it hits the land surface, into streamflow generally takes place over an area of land along the river channel that extends from the channel banks to the nearest divide. Thus each channel segment of a river system can be visualized as lying between two strips of riparian land on either side that feed water into it. While the mechanisms involved in the transformation from precipitation to stream flow depend on many factors, an important one to consider is the relative size of the river segment within the river and tributary system of the basin.

In geomorphology, it is customary to classify stream channels in a hierarchy of orders, in which the order of a stream depends on the number of upstream tributaries or bifurcations. Horton (1932; 1945) was probably the first to propose a downstream-moving ordering procedure. In this system, tributaries without branches are called first-order streams; the branches that receive only first-order streams are designated second-order streams, and those that receive one or more second-order and also

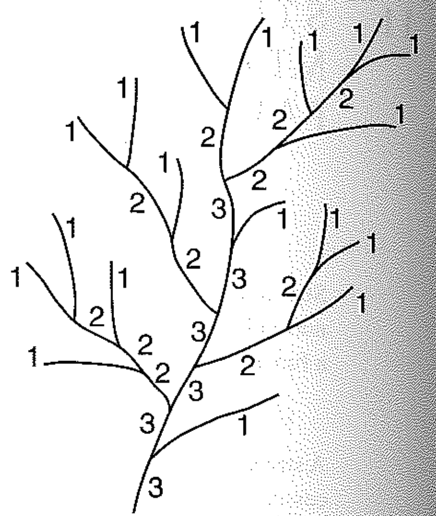


Fig. 11.1 **The order numbers of river channel segments in a natural basin drainage network according to the Horton–Strahler method.**

first-order streams are considered third-order streams, and so on. The definition of first- and second-order streams is clear and unambiguous in Horton's procedure, but the definition of third- and higher-order streams required some subjective decisions. To avoid these and to ensure that only one stream would bear the highest-order number in the basin, Strahler (1952) adjusted the procedure by stipulating that third-order streams can only be formed by the joining of any two second-order segments, and so on. The Horton–Strahler method, as it is now called, is illustrated in Figure 11.1; in this example, there are 18 first-order streams, five second-order streams, and one third-order stream.

Larger-order river channels usually do not receive much water locally from the riparian surfaces along their banks, but they receive most of their water from upstream through lower-order streams. The catchments that are drained by lower-order streams with no or very few tributaries can be called *headwater* basins, *source area* watersheds, or also *upland* watersheds. Because they feed into channels of progressively higher order, these lower-order catchments are crucial for a better understanding of runoff mechanisms in larger basins as well. An important feature for the analysis of runoff from such headwater catchments is that lower-order river channels tend to have relatively short residence times; thus any storm runoff hydrograph from a source area watershed is affected primarily by the nature of the soil mantle areas surrounding the stream and very little by the nature of the stream itself. Further downstream, however, as more and more tributaries join, the shape of the hydrograph evolves, and it will increasingly reflect the hydraulic characteristics of the channel network. The flow mechanisms in riparian areas and headwater basins, a topic often referred to also as *hillslope hydrology*, have been the subject of intense research in the past few decades. A knowledge of these mechanisms and of their interactions is not only essential to describe streamflow generation, but it is also the key to a better understanding of solute transport in the human environment and of the evolution of landforms and erosion.

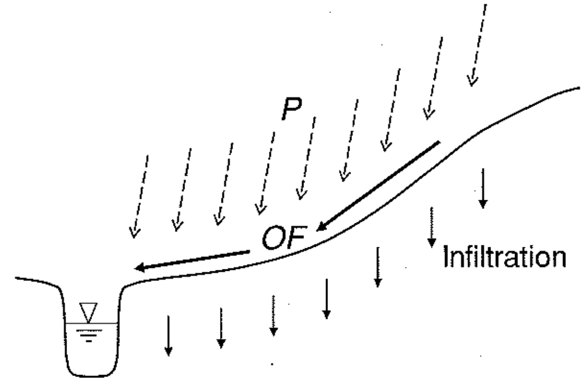


Fig. 11.2 **Illustration of the overland flow (OF) mechanism as infiltration excess. The precipitation rate *P* exceeds infiltration capacity, and the water table is at the ground surface.**

11.2 STORM RUNOFF MECHANISMS IN RIPARIAN AREAS

11.2.1 Overland flow

Infiltration excess overland flow

This type of flow occurs when the rainfall rate is larger than the infiltration capacity, so that there is an excess which runs off over the surface. Although this flow generation concept is sometimes associated with the name of Horton (1933), it goes back much earlier. It was already the basis of the well-known rational method, introduced 150 years ago by Mulvany (1850), and of the various runoff routing procedures subsequently derived from it by Hawken and Ross (1921) and others (see also Dooge, 1957; 1973). It is also implicit in the unit hydrograph, as originally proposed by Sherman (1932a; b). In these and other early studies concerned with maximal rates of runoff in problems of flooding and erosion, it was assumed that the infiltration rate is smaller than the precipitation rate over the entire catchment. In the rational method, the infiltration is taken as a fraction of the precipitation, whereas in the unit hydrograph approach and in Horton's work, the infiltration capacity or a related index is subtracted from the precipitation. Thus it was assumed that the infiltrated water is "lost" and that virtually all stormflow results from the overland flow of the precipitation excess (see Figure 11.2). In the prediction of extreme flows for design purposes in disaster situations, this assumption of overland flow was not unreasonable.

It is now understood that overland flow is not a universally occurring phenomenon, that in many situations it may not occur at all, and that its prevalence depends on the nature of the catchment and of the intensity of the precipitation. But it can be expected to be the main mechanism in catchments with relatively impermeable surfaces, and with only a thin soil layer; such surfaces cover mostly urban environments, factory and farm yards and other trampled soil areas, and rocky and stony areas with little or no soil or vegetation, as seen in arid and desert environments. Thus it occurs most frequently in areas where people live and work and in denuded arid regions. It can also occur on other more permeable surfaces, provided the rainfall is sufficiently intense. For instance, in a study of a 20 ha first-order agricultural catchment with steep slopes in semi-arid Shanxi (China), Zhu *et al.* (1997) reported that most storms generate no overland flow. However,

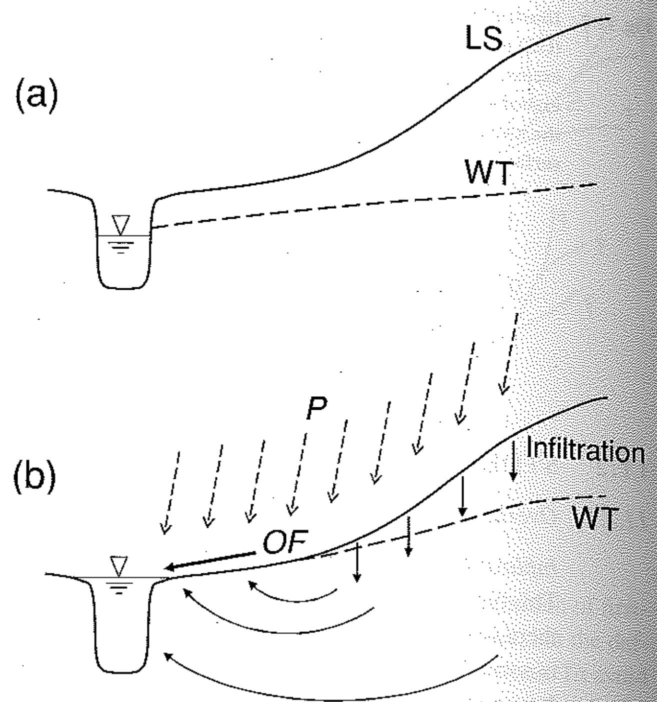


Fig. 11.3 Schematic illustration of the overland flow (OF) mechanism as saturation excess: (a) the position of the water table (WT) prior to the onset of precipitation and (b) during the precipitation event. The precipitation rate *P* is smaller than the infiltration capacity over the unsaturated portion of the land surface; overland flow takes place where the water table has risen to the ground surface.

in 8% of the precipitation events infiltration excess overland flow was the predominant runoff process; rainfall intensity, rather than rainfall amount, was the decisive factor for its occurrence, although soil surface crusting also played a role. Occurrence and yield of overland flow varied spatially on account of the variability of the infiltration capacity.

In general, infiltration excess overland flow appears to be rare in natural basins covered with a thriving vegetation in more humid climates.

Saturation excess overland flow

This type of surface runoff occurs over land surfaces that are saturated by emerging subsurface outflow from below and perched water tables, regardless of the intensity of the rainfall (or snowmelt) (see Figure 11.3). It is a rapid and almost immediate transport mechanism to the stream channel, for the seepage outflow water and for the rainwater falling (or snow melting) on such areas. It usually takes place in conjunction with subsurface flow to the channel, but the relative magnitudes of surface and subsurface flows into the channel depend largely on the nature of the catchment and the precipitation. It is most often observed over limited areas in the immediate vicinity of the river channel where downslope subsurface flows emerge, and in wetlands, where the water table can rise rapidly to the surface; but it can also occur higher up in slope hollows, where elevation contours display strong curvature, thus forcing convergence of the flow paths. Outside of these saturated areas all the precipitation and other input can generally enter the soil surface.

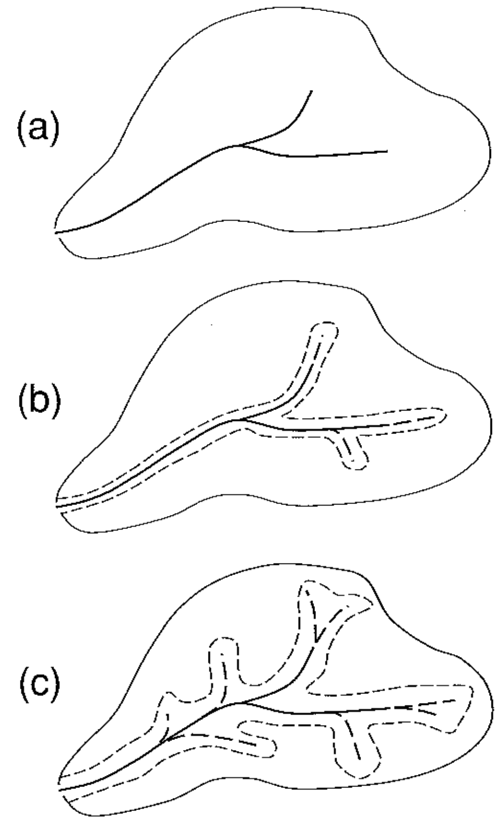


Fig. 11.4 Schematic plan view of a second-order catchment illustrating the extent of the variable source areas (inside the dashed line) on which overland flow takes place: (a) under drought flow conditions; (b) and (c) after the onset of precipitation. The stream channels and the saturated areas near the stream channels expand as the precipitation continues.

For instance, as early as 1961, US Forest Service hydrologists (Hewlett, 1974; Hewlett and Hibbert, 1967) reported that in forested hilly catchments in the Coweeta section in the southern Appalachians of North Carolina, the streamflow hydrograph rises as a result of precipitation on the channel itself and as a result of the expansion of these saturated areas in its immediate vicinity. The expanding and shrinking areas are often referred to as *variable source areas* (see Figure 11.4). On the basis of hill slope measurements in Vermont, Dunne and Black (1970a; b) also concluded that the stormflow originated from surface flow on limited areas along the stream channel. However, their interpretation of the mechanism was that this surface runoff was not fed significantly by subsurface outflow, but resulted mostly from rainfall on the expanding streamside areas; the role of the subsurface flow was mainly to control the expansion and subsequent contraction of the source areas.

But saturation excess overland flow does not always occur in the immediate vicinity of the stream. In a tropical rainforest in northeast Queensland, Bonnell and Gilmour (1978) and Elsenbeer *et al.* (1995a) observed that high intensity rainfalls generate widespread perched water table conditions close to the soil surface, which emerge easily; this results in saturation excess overland flow accompanied by subsurface flow within the top 20 cm. Evidence for this was taken to be the presence of pre-event water in the streamflow, that is water which was present in the soil profile prior to the rainfall event; if infiltration excess overland flow had been the only mechanism, all the storm runoff would have been event water, that is water furnished by the rainfall event. The ratio of event to pre-event water in the streamflow was found to depend on the rainfall duration and intensity. Because overland flow was so widespread, they concluded that in this type of tropical rainforest the variable source area concept does not apply. Elsenbeer (2001) subsequently surmised that overland flow may be a common flowpath in tropical rain forest catchments with "acrisol" profiles; these are soils, in which the clay content increases with depth, resulting in a decreasing hydraulic conductivity.

11.2.2 *Subsurface stormflow*

In many catchments under natural conditions infiltration is never exceeded, and the precipitation and other input can readily enter into the ground surface; thus the subsequent flow to the stream channel takes place below the surface, presumably through the soil mantle of the catchment. Lowdermilk (1934) and Hursh (1936) appear to have been among the first to propose subsurface flow as the main streamflow generation mechanism in forested hill slopes (see also Hewlett, 1974). It was later confirmed in several experimental investigations that subsurface flow can even be the only mechanism under certain conditions (see Roessel, 1950; Hewlett and Hibbert, 1963; Whipkey, 1965; Weyman, 1970).

The notion that subsurface flow is an important, and sometimes the only process of water transmission, was resisted by many on the grounds that porous media flow is generally much too slow compared with overland flow to be able to produce the observed streamflows. One early explanation of this paradox was suggested by Hursh (1944), who assumed that the transport takes place through secondary porosity of particle aggregates, forming a three-dimensional lattice pattern, and through hydraulic pathways consisting of dead root channels and animal burrows (see also Section 8.3.1). At the time, this possibility of macropore flow and piping seems to have been largely dismissed as unrealistic by experimentalists and mostly ignored by modelers. However, subsequent experimental work in the field, some of it with chemical and isotopic tracers, has produced ample and incontrovertible evidence not only for macropore flow and its importance, but for several other mechanisms enhancing subsurface flow as well. These are considered more closely in what follows.

Macropores and other preferential flow paths

The concept of preferential flow paths or macropores is an old one; "little channels" and "light soil, mixed with pebbles and roots of trees" were invoked as early as the 1680s by Mariotte to explain infiltration and to refute the claims of Seneca and Perrault that rain water cannot possibly penetrate the soil to be the source of springs. In general, macropores can be defined as secondary, often pipe-like structures of the soil matrix, that are the remains of purely physical processes, such as erosion initiated by desiccation cracking, and different forms of biological activity, such as decaying plant root channels

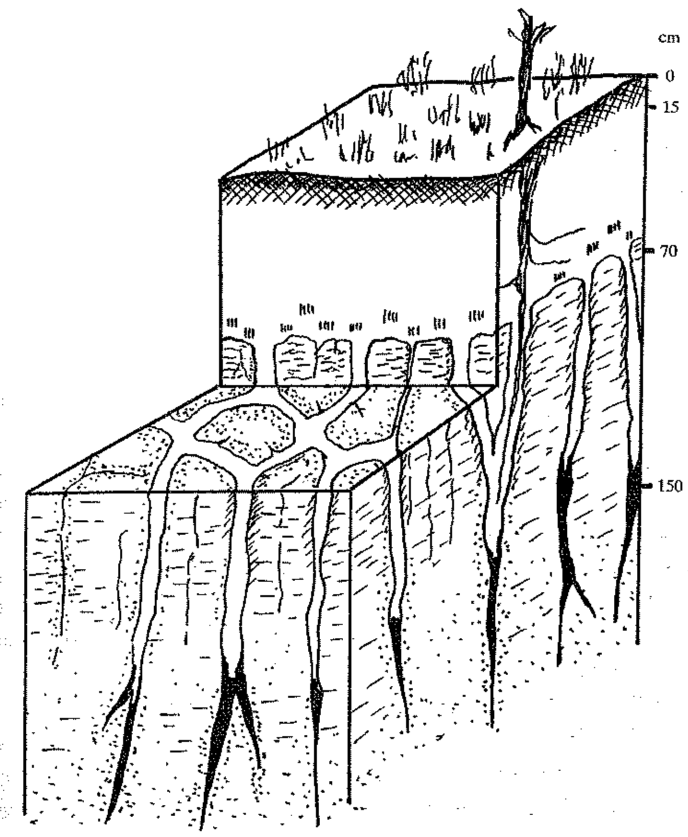


Fig. 11.5 Schematic representation of a soil profile with a fragipan horizon. (From Smalley and Davin, 1982, after Van Vliet and Langohr, 1981.)

and animal burrows of various sizes. Reviews on the subject have been presented by Jones (1971) and Beven and Germann (1982). Because soil drying and biological activity tend to take place near the ground surface, pipes and macropores are usually most abundant in the top soil layers and they tend to become less frequent with depth. Such structures are usually obvious features of soil profiles at banks and road cuts.

In addition to these macropores, different other types of preferential flow paths have been observed, which may also have ramifications for the relative transport of *pre-event* and *event* water to the stream channel. Recall that pre-event water, also called *old* water, is the water present in the soil mantle before the onset of the precipitation, whereas the event water, often called *new* water, is the water resulting from the precipitation. In one type of preferential flow, the paths can often be observed at the surface of clayey and loamy soils as cracks or fractures resulting from the shrinkage of the clay particles during drying episodes. At least during the initial stages of a precipitation event, before clay swelling closes them again, such cracks can facilitate the downward water movement in the profile. A somewhat related type of preferential flow has been observed in fissured fragipan horizons (Parlange *et al.*, 1989), as illustrated in Figure 11.5. A fragipan is typically a loamy clay layer with very low conductivity and higher bulk density than the overlying layers. However, in some cases during their evolution, fragipan horizons became fractured into a polygonal columnar structure with a network of interconnected vertical fissures, again, as a result of shrinkage of the clay particles; these cracks are then believed to have been filled with more permeable soil material from above, greatly facilitating water transport. The cracks are typically 10–20 cm wide. In another type of preferential flow, the paths are initially established as instabilities or *fingers* at infiltrating wetting fronts in coarse soils, when the infiltration rate is smaller than the saturated conductivity. A crucial point, however, is that, once established, these paths usually become permanent features of the profile, each time the soil is being rewetted (Glass *et al.*, 1989); exceptions may occur when the soil has undergone complete drying out or complete saturation, both of which are rare if not unlikely in nature. Figure 11.6 shows an example of the initial growth of fingers observed in the laboratory. Such fingers are not so obvious in the soil profile, but they become visible with dyes or other tracers. Other aspects of the nature and origin of this type of preferential paths have been clarified (see Selker *et al.*, 1992; Liu *et al.*, 1994a; b).

Although the existence of macropores has been known for a long time, the precise nature of their contribution to the streamflow generation processes has been emerging only gradually. A few examples follow of investigations in which macropores were observed to play a major role.

In a small (0.022 km^2) basin in east-central Honshu, Tanaka *et al.* (1981; 1988) observed that more than 90% of the storm runoff came from below the ground surface mainly through pipe flow; some saturation overland flow occurred over the gentler slopes ($S_0 \cong 0.12$) of the valley floor, when the rainfall exceeded 50 mm; the saturated area varied somewhat in location and extent from storm to storm, but it never occupied more than 4.5% of the total area (see Figures 11.7 and 11.8). No overland flow was ever observed on the steep ($S_0 \cong 0.50$) hillsides.

In a 0.47 ha forested catchment in Tennessee, Wilson *et al.* (1991) found that the initial subsurface stormflow water in moderate to high intensity events consisted mainly (>70%) of new, i.e. event water; they concluded from this that it had bypassed the unsaturated soil matrix, in which the pre-event water was stored, via macropores without ever reaching the water table. Later on, however, as the flow continued, the fraction of old water gradually increased.

In a catchment under pasture in southern Australia, Smettem *et al.* (1991) and Leaney *et al.* (1993) observed that winter stormflow reaches the channel mainly through macropores, bypassing the soil matrix, and creating perched water table conditions immediately around these pores. In summer, however, overland flow was found to be dominant; they did not observe evidence of partial area sources, as only a negligible fraction of the catchment was occupied by wetland.

On a steep forested hillslope with cedar and cypress in Ibaraki in east-central Honshu, Tsuboyama *et al.* (1994) observed a dynamic system of macropores, which expanded and conducted increasing amounts of water as antecedent conditions became wetter. Continued studies on that same catchment (Noguchi *et al.* 1999; Sidle *et al.*, 2001) led

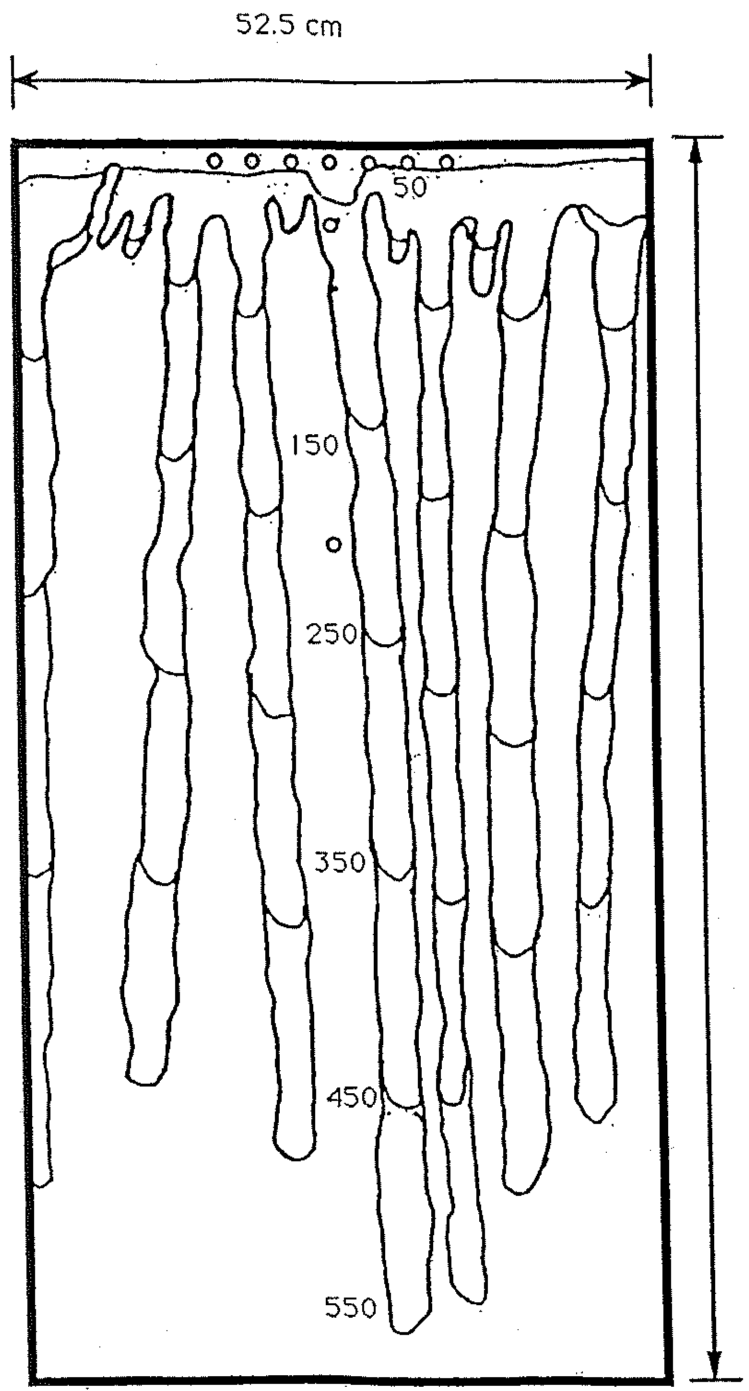


Fig. 11.6 **Typical development of an unstable wetting front resulting in a persistent fingered flow pattern; the round holes indicate the positions of the tensiometers in the two-dimensional sand-filled chamber to monitor the water pressures during the experiment, and the numbers indicate time (s) after the start of the infiltration. (From Selker *et al.*, 1992.)**

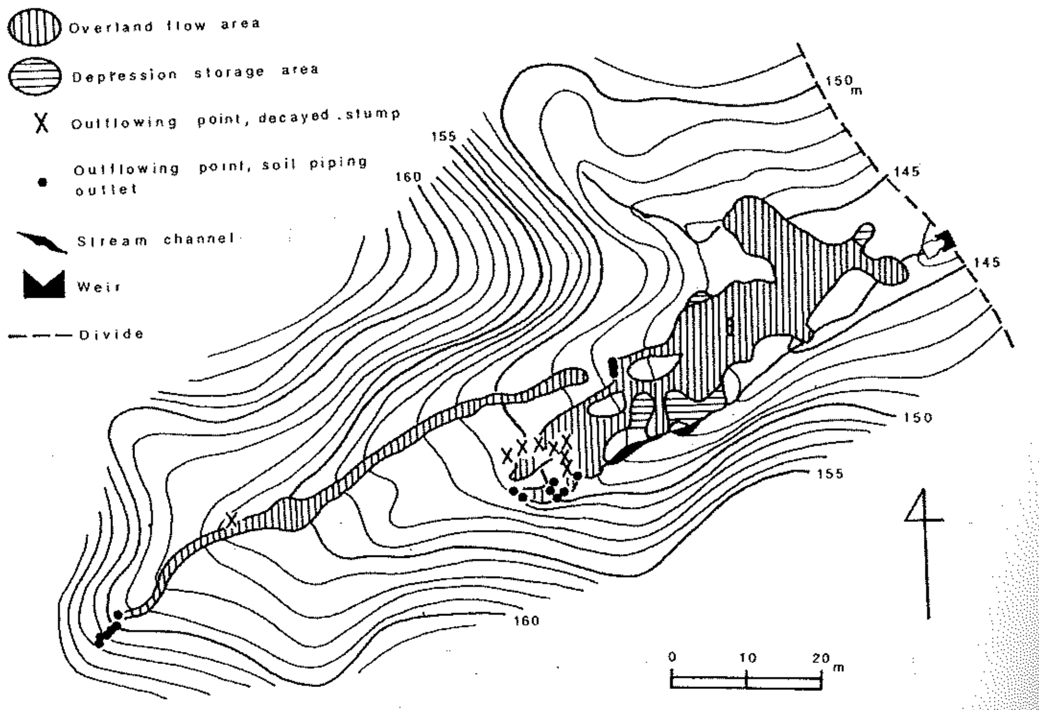


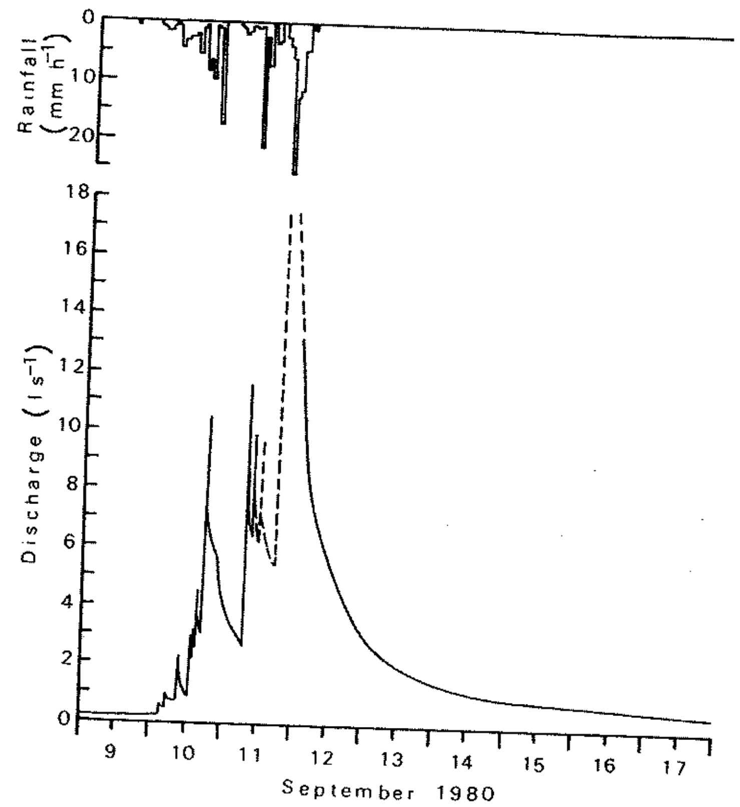
Fig. 11.7 **Maximal extent of the saturated areas and distribution of subsurface outflow points at the peak discharge rate of a 195 mm storm in September 1980 on a steep 0.022 km^2 catchment within the source region of the Tama River. The saturated area occupied roughly 3.3% of the basin area and the area shown represents roughly one quarter of the basin area. (From Tanaka *et al.*, 1981.)**

to the more specific view that, while individual macropore segments are usually shorter than 0.5 m, they tend to self-organize, as wetness increases, into larger flow systems with such preferential flow connections between them as buried pockets of organic material and loose soil, small depressions of bedrock substrate, and fractures in the weathered bedrock.

Chemical analysis of measurements on a 0.75 ha forested first-order catchment in the sub-Andean foreland basin of Peru by Elsenbeer *et al.* (1995b) indicated that the stormflow response is dominated by event water. This water traveled to the stream channel as a combination of overland flow and through pipes. Some pipe flow reached the stream directly, but some emerged to the surface before reaching the stream. The overland flow was thus generated by emerging pipe flow and directly by the rain. This made them observe that, from the perspective of the catchment, the distinction between pipe flow and overland flow is meaningless, as both mechanisms produced event water.

From observations in a semiarid pine forest in New Mexico, Newman *et al.* (1998) concluded that most of the lateral subsurface flow takes place in the B horizon through macropores. Thus throughout most of the year, the soil profile behaves like a two-domain

Fig. 11.8 **The precipitation and runoff hydrograph of the 195 mm storm event on the catchment shown in Figure 11.7. (From Tanaka *et al.*, 1981.)**



system; this consists of a macropore domain, which provides rapid subsurface flow that is not in equilibrium with the soil matrix, and of a matrix domain, in which the transport is very slow and in which evaporative processes cause major water losses and increased salinity (see Figure 11.9). Variations in the ratio of old to new water in the runoff were seen to depend mainly on the size of the precipitation event; macropores can conduct the flow directly or they may also feed shallow perched saturated zones overlying low permeability bedrock. Whenever the entire profile is fully saturated, as during snowmelt episodes, the two domains are connected, and large subsurface flow rates are produced.

In the above studies it was shown how subsurface flow through macropores and other preferential flow paths can play a major role in storm runoff generation. However, the specific interpretations of the measurements, especially on the relative roles of old and new water in this process, differed somewhat, and in some cases they were contradictory. Although this is largely the result of the wide variety in catchments that were being studied, it is no doubt also related to the differences in experimental techniques used in these studies. This was brought out, for example, in the long-term observations, carried out on the steep mixed evergreen forest catchments (1.63–8.26 ha) in a humid climate (2600 mm y^{-1}) at Maimai in New Zealand; a succession of detailed studies has

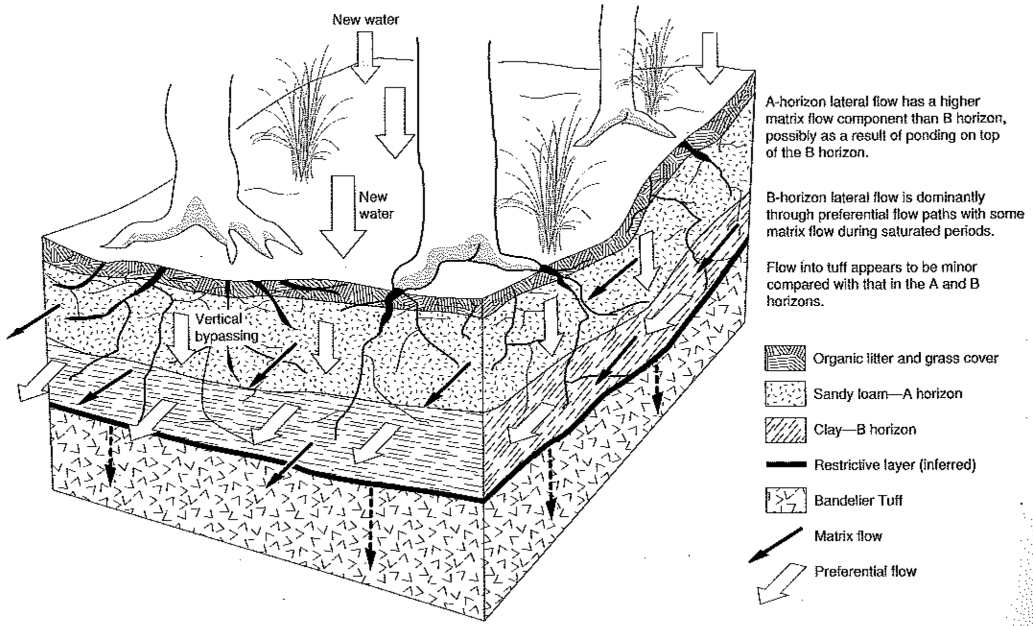


Fig. 11.9 Perceptual flow mechanisms in a semiarid forested slope in New Mexico. The lateral matrix flow in the A horizon is larger than that in the B horizon, possibly as a result of some ponding on top of the B horizon; in the B horizon the flow takes place mainly through preferential flow paths, with some matrix flow and leakage into the underlying tuff. (From Newman *et al.*, 1998.)

illustrated how such interpretations can evolve over time, as more and better measurement techniques are brought to bear on the analysis (McGlynn *et al.*, 2002). In the early studies by Mosley (1979) it was concluded from local flow and dye tracer measurements in pits that macropore flow of mostly new water, in storms of moderate to large intensity, can bypass the soil matrix, where the pre-event water is normally stored, and is capable of generating the channel stormflow. On the basis of subsequent investigations with electrical conductivity and natural tracers, Pearce *et al.* (1986) and Sklash *et al.* (1986) arrived at a different conclusion; they deduced from the measurements that it was mainly old water throughflow that was responsible for hydrograph generation, and that the flow of new water above the ground surface or below it through the soil matrix or through macropores could not explain the streamflow response. To resolve these discrepancies, a third set of studies was carried out by McDonnell (1990; McDonnell *et al.*, 1991a) in which a chemical tracer analysis was supplemented with soil water pressure observations by means of tensiometers installed in near-stream, mid-hollow and upslope positions. It was observed that the soil water pressure response was dependent on storm magnitude, intensity and antecedent water content. For storm events producing peak runoff less than 2 mm h^{-1}, the water appeared to infiltrate downward as a wetting front in the soil matrix without appreciable macropore bypass flow; no water table developed along the slope and the streamflow consisted of old water issuing mainly from the near stream valley bottom groundwater. For events with peak storm runoff in excess of 2 mm h^{-1},

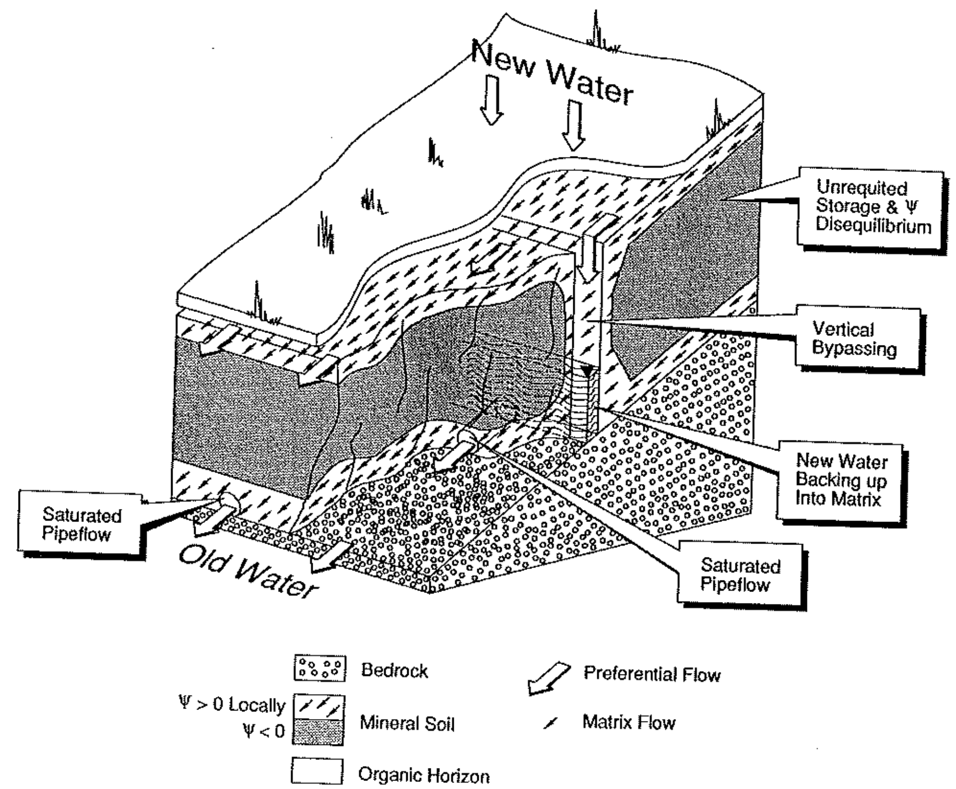


Fig. 11.10 Perceptual runoff production mechanisms in a midslope hollow of a humid catchment in New Zealand. As shown, the precipitation rate (*P*) exceeds the hydraulic conductivity (k_0) of the mineral soil, and moves down through vertical cracks. The invading new water perches at the soil–bedrock interface, and backs up into the newly saturated soil matrix, where it mixes with the much larger volume of stored old water. Once free water (with positive pore water pressures) exists, the larger pipes in the lower soil zones quickly dissipate transient water tables laterally downslope, producing a rapid throughflow response of well-mixed, albeit mainly pre-event water. (From McDonnell, 1990.)

the lower soil horizons along the slope responded almost instantaneously, indicating a rapid macropore flow, as Mosley (1979) had already surmised. The predominance of old water in the streamflow runoff was explained by McDonnell (1990) by the fact that the rapid flow of new rainwater through downward crack macropores backs up into the soil matrix at the soil–bedrock interface, which is still dry; this rapidly causes saturated conditions, and results in the emergence of well-mixed old water from the matrix into lateral pipe macropores and rapid downslope transport (see Figure 11.10). In a fourth set of experiments, Woods and Rowe (1996) dug a trench 60 m long and 1.5 m deep along the toe of a hillslope hollow, with 30 subsurface flow collection points along its length. The outflow from the hillslope was found to be very variable; this led to the conclusion that outflow data from single hillslope throughflow pits should not be extrapolated to an entire hillslope and further (Woods *et al.*, 1997) that this variability depends on wetness and surface topography. The latter conclusion was refuted by McDonnell and associates

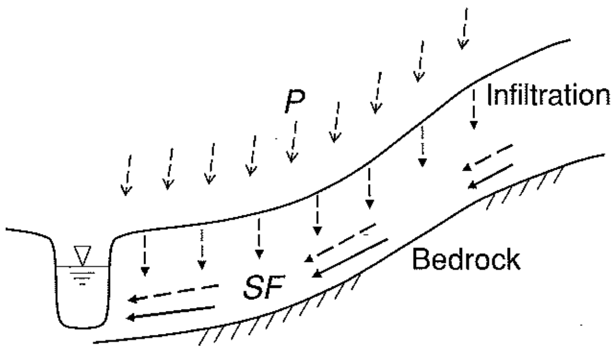


Fig. 11.11 Schematic illustration of the rapid subsurface storm flow (SF) through various types of preferential flowpaths, pipes and macropores. The relative amounts of new (dashed arrows) and old water (solid arrows) in the mixing process depend mainly on the precipitation intensity and on the pre-storm soil moisture conditions.

(McGlynn *et al.*, 2002) on the basis of a fifth set of hillslope-scale tracer measurements with bromide at the same catchment. The main conclusion from that study was that it is not the surface topography, but rather the spatial pattern of the bedrock topography, with local preferential flow and mobile and immobile regions, conditioned by small local depressions in the bedrock, which controls the tracer outflow variability; tracer material and old water may remain trapped temporarily in such depressions and become mobilized only by a new storm event.

From measurements in a forested Canadian Shield basin in Ontario, Peters *et al.* (1995) concluded that preferential flow channels brought the water vertically down, after which it flowed laterally over the bedrock and that practically all the lateral flow occurred within a thin weathered zone near the soil–bedrock interface. The conductivity of this preferential flow layer appeared to be so large that some of the fast flows and peak runoff were suspected to be of the non-Darcy type. The storm runoff in the channel consisted of a mixture of event and pre-event water. This was interpreted to show that the fast infiltration of the event water caused saturated soil conditions above the bedrock, which in turn resulted in the downhill flow of both event and pre-event water; moreover, during the hillslope transport, there was ample opportunity for interaction between the event runoff water and the soil matrix.

In summary, the subsurface stormflow, observed in several of the hillslope experiments reviewed here, exhibited the common feature of unimpeded entry by new water from rainfall into the soil, followed immediately by rapid downslope flow through preferential paths, pipes and other macropores; this flow involved mixing with the old water already present in the soil profile (see Figure 11.11), to varying degrees depending on the intensity of the rain and on the initial moisture status of the soil mantle.

Throughflow in a shallow permeable layer

In many catchments covered with natural vegetation the soil mantle has a relatively permeable top layer consisting of organic debris and mineral soil with high organic content; typically, this layer has a thickness of only a few tens of centimeters and its bottom interface is characterized by an abrupt decline in hydraulic conductivity in the underlying mineral soil. Thus infiltrating rainwater tends to flow and build up along this interface and develop a perched water table and fully saturated conditions, although deeper layers may remain partially saturated. In several experimental studies it has been observed that such layers can be effective enough to be a major, and sometimes even the main transport medium for stormflow. As noted above, this type of flow was observed to occur by Bonnell and Gilmour (1978), in conjunction with saturation overland flow, in a catchment in Queensland. The chemical signatures of hillslope waters in a catchment in Wales made also Chappell *et al.* (1990) conclude that this can indeed be the dominant mechanism for water and ion transport to lower near-stream riparian zones. Similarly, Jenkins *et al.* (1994) used natural tracers to characterize rain water, soil water and ground water in a moorland catchment in northeast Scotland. The interface between mineral soil layers and the upper organic layers of peaty podsol were identified as preferential pathways. Flow of water in this upper layer was observed to be triggered nearly instantaneously by the onset of the rain, and also to stop nearly as suddenly as the rain ceased; the water in this layer had a chemistry very similar to that of the rain. In the runoff hydrograph, the peak flow was found to be dominated by rain and soil water, whereas the recession part was dominated by pre-event groundwater.

Although they did not consider this type of flow as being representative of the entire catchment, McDonnell *et al.* (1991b) did observe it on small portions of the Maimai catchments in New Zealand. During a rain storm event of some 47 mm and with the soil water suctions initially ranging between $H = 60$ and 150 cm water column, most of the water was seen to flow out from the organic soil layer perched on the mineral soil profile; all the while, the lower soil profile remained only partly saturated. More recently, from an experimental study on seven nested (from 8 to 161 ha) forest catchments in the Catskill Mountains of New York, Brown *et al.* (1999) concluded that a large fraction of the rapid delivery to the stream took place through this same mechanism. Event water appeared to be most prevalent in the stormflow especially during dry conditions, with relative contributions between 50% and 62% near peak flow.

Wavelike mobilization of the water table

As illustrated in Figures 8.5, 8.6 and 8.7, for most soils within the nearly saturated capillary fringe, a small change in water content can result in a relatively large change in pore water pressure. This has led to the view that the addition of a very small amount of water to a relatively moist soil can raise the water table rapidly, almost as a pressure wave type of propagation, to produce a saturated soil profile. Wherever the profile becomes fully saturated this way, subsurface flow may emerge and saturation excess overland flow is also bound to occur. This type of water table rise may be especially fast in the lower parts of the hillslope and may result in the build up of an emerging groundwater mound, exhibiting greatly increased hydraulic gradients and groundwater discharge to the channel, and forming a partial or variable source area producing saturation excess overland flow as well. Thus the phenomenon is not unlike that depicted in Figure 11.3, except that here the water table rise is presumed to involve very little actual water movement.

Phenomena interpreted to be the result of this type of mechanism were observed, for example, in a swampy area by Novakowski and Gillham (1988) and in a grass-covered low relief basin by Abdul and Gillham (1989), both in Ontario. In these studies, the rise of the water table was most pronounced in the near-stream areas. The mechanism has also been inferred to occur in more rugged terrain. During a sprinkler irrigation experiment on a very steep (43%) forested hillslope near Coos Bay in Oregon, Torres *et al.* (1998) applied a sudden input spike, after the system had been driven to a steady state flow and the soil water pressures were mostly between 0 and −10 cm. They supposed that the timing and magnitude of the pore water pressure and of the discharge rate response to this sudden input were much faster than could be expected from advective water movement, and concluded that the fast response was triggered by a pressure wave moving undetected through the unsaturated zone; thus a small amount of rain on a wet soil profile can supposedly result in a rapid rise in the saturated zone, with a relatively slight increase in hydraulic gradient and a large increase in hydraulic conductivity. They also observed some preferential flow, but they felt that in this particular soil, this effect was minor compared to that of the soil water retention characteristics.

The concept that suction-saturated capillary fringe water can be easily converted into water below the water table, that is from a negative to a positive pressure, by a relatively small amount of rain, is undoubtedly realistic. Clearly, only a little additional water is required to mobilize the soil water, when the soil is already close to saturation. But the importance of this mechanism should be kept in perspective. For example, it can only be expected to be effective when the pore water pressure in the top layers of the soil is arrived at during a drainage phase and not during a wetting phase; as illustrated in Figures 8.14, 8.18 and 8.19, the capillary fringe is usually much smaller in the wetting cycle. Similarly, in the absence of any macropores or pipes, the water table (i.e. the locus of atmospheric- or zero-pore water pressure) can only be expected to move rapidly down a steep slope, if it is already close to the surface. As illustrated in Figure 10.12, the drainable porosity n_e is smaller when the water table is closer to the surface, that is when $(-p_w)_{max}$ is smaller. While not a perfect representation, hydraulic groundwater theory, as formulated by the Boussinesq equation (10.29) and its linearized form (10.134), is also fully consistent with this. This can be seen by considering the advectivity in Equation (10.29) (also (10.136)); rewritten here for convenience

$$c_h = -\frac{k_0 \sin\alpha}{n_e} \tag{11.2}$$

it describes the speed of propagation of a given water table height η (or of a disturbance of the water table resulting from rainfall) down the slope. This shows that, in the absence of preferential flow paths, large values of c_h can result only when the drainable porosity is small. As seen in Equation (10.151), this is equally consistent with the kinematic wave approximation.

Capillarity induced flow enhancement has also been linked to soil stratification. In situations where a fine-textured soil layer overlies a more coarse-grained material, the interface between the two layers can develop into a capillary barrier (Ross, 1990; Steenhuis *et al.*, 1991). On account of the different soil water characteristics in the two layers, for a given water pressure at the interface, at equilibrium the soil water content in the upper layer is normally larger than in the lower coarse-textured layer. As a result, the hydraulic conductivity in the upper layer may be considerably larger than in the lower layer; this is illustrated in Figure 8.26. In such a case infiltrating rain water will not readily enter into this lower layer but will tend to be diverted laterally and may cause a rapid rise in water table further down the slope if the water in the upper layer is already close to suction-saturated. Field observations within the source region of the Tama River in east-central Honshu by Marui (1991; Tanaka, 1996) on a hillslope unit, characterized by a 4 m thick fine-grained loam layer underlain by 15 m thick gravel layer, were consistent with this sequence of events. He observed a large-scale groundwater ridge along the steep hillslope. In addition, the air in the underlying partly saturated gravel seemed to be confined by the surrounding groundwater body, and by the saturated zone in the loam layer. In a separate study, Onodera (1991; Tanaka, 1996) inferred that the resulting air pressure increase may have led to increased groundwater outflow at the slope surface.

In conclusion, it stands to reason, that mechanisms related to capillarity can lead to so-called groundwater "ridging" not only in riparian areas, but also along hillslopes, wherever the capillary fringe is already close to the ground surface. However, until now no experiments have demonstrated that by itself this type of phenomenon is related to the hydrograph; thus, whether or not this mechanism can explain large subsurface stormflows remains to be answered.

11.3 SUMMARY OF MECHANISMS AND PARAMETERIZATION OPTIONS

11.3.1 *General considerations*

The brief review in Section 11.2 has shown that on the Earth's land surfaces one can encounter a bewildering range of hydrologic, climatic, topographic and soil conditions, which will favor widely different storm generation mechanisms. These mechanisms can be overland flow due to infiltration excess precipitation, or to saturation excess near the soil surface, resulting either from return outflow from the subsurface, or from rapidly mobilized capillary fringe water in the soil profile to full saturation. On steep slopes overland flow is more likely on converging sections in hollows. The mechanisms can also be subsurface flow of water in a number of different ways. Especially during large rainfall events, this can involve different types of macropores and preferential flow paths, namely as vertical bypass flow to some depth, and then as lateral flow through pipes or through a shallow porous soil layer with high organic content or at the soil bedrock interface. At the same time a slower and less localized throughflow takes place in the soil matrix. Several of these mechanisms have been found to be more than adequate to produce high-intensity runoff events. It is also striking that these mechanisms are not mutually exclusive and that in many situations they coexist and operate interactively in the production of streamflow; their relative importance then depends on the prevailing conditions, such as initial moisture conditions in the catchment and the magnitude of the precipitation.

In some cases the coexistence of different mechanisms can give rise to some unusual phenomena. For instance, under low initial conditions in a watershed in central Côte d'Ivoire, Masiyandima *et al.* (2003) observed double-peaked hydrographs resulting from the same rainfall burst; the first peak, which occurred while it rained, was produced by the rainfall on the saturated valley bottom; the second peak, which came minutes to hours after the first, resulted from the rain that had fallen on the area surrounding the valley bottom and that had traveled to the stream channel by subsurface flow.

All this underscores again the extreme complexity of the stream generation process. These observations suggest that a single unifying runoff model may not be possible nor even desirable, and they have profound implications for the development of modeling strategies for predictive purposes in applied hydrology.

Identification of major mechanisms

In order to keep the formulation sufficiently simple and parsimonious, it may be necessary to identify and include only the dominant mechanisms for any given set of conditions, and to accept some inevitable uncertainty resulting from the omission of the remaining minor mechanisms. On the basis of a knowledge of these local conditions, the analyst must then decide which mechanisms are the major ones that must be considered to represent a particular catchment. The insight gained by the recent field observations can also give some guidance in this. For instance, different kinds of subsurface flow can be assumed to dominate the runoff process in humid areas with an active vegetation. Well-developed mineral soils undoubtedly favor the development of preferential flow paths, whereas thin porous soils with organic litter probably lead to shallow lateral flow of the perched water above the less permeable soil or bedrock. Wetland areas near the stream may allow rapid mobilization or ridging of the water table, and the development of partial and variable source areas, on which saturation excess overland flow can take place. Infiltration excess overland flow will be prevalent during large precipitation events on unvegetated surfaces in arid regions and in areas subject to intense human activities.

Objectives of the analysis

This wide variety in possible mechanisms also means that, in the development of modeling strategies for engineering and other applied purposes, for a given catchment it may be advisable to adopt different formulations depending on the objectives of the analysis. For example, the prediction of disastrous flash flooding, under extreme precipitation conditions, may require an entirely different approach from those needed to describe solute transport and water quality in the environment, to analyze possible climate change scenarios under more normal flow regimes, or to assess the potential for erosion or landslides. For flood prediction, mainly the flows at a certain point along the river may be of interest; for climate change scenarios, surface–atmosphere interactions are of paramount importance; and for water quality purposes it may be crucial to know the pathways, in order to determine the fate and transport of admixtures and water pollutants; finally, erosion and landslide hazards tend to be related to the pore water pressure distribution and the local flow velocities.

Appropriate parameter values

But even for the same formulation, it may also be necessary to adopt different parameter values depending on the flow regime. The formulation of river flow usually requires different values of the roughness parameter in the Gauckler–Manning or Chézy equations, depending on whether it is low flow within the regular channel or high flow with flooding outside the banks. Similarly, in the description of hillslope outflow by some of the subsurface parameterizations of Chapter 10, the appropriate values of the effective hydraulic conductivity k_0 and of the thickness of the flow region η_0, used to represent stormflow conditions with active macropores, will be considerably larger than those appropriate for conditions of baseflow, after the water tables have subsided and many of the macropores in the upper soil layers have emptied and are no longer active. Actually, because of the high flow velocities, subsurface stormflows may not be of the Darcy type, and it may be necessary to use Forchheimer's Equation (8.34) with an additional transmission parameter beside the hydraulic conductivity.

Ultimately, the performance, in a general sense, of any kind of parameterization and of the resulting model, has to be judged on the basis of its ability to simulate or replicate observations of the variables of interest. As mentioned in Chapter 1, parsimony and robustness are important additional considerations. Different aspects of the modeling issue have been treated by Klemes (1986), Morton (1993), and Woolhiser (1996), among others.

11.3.2 *How to put it all together? Distributed versus lumped approach*

As already explained in Chapter 1, scale is the appropriate criterion to classify the different methodologies. Accordingly, one can distinguish two general classes of models that have been used in the past to simulate streamflow generation. In the *distributed* models, also called *runoff routing* models, the computational scales are much smaller than the flow domain characterizing the catchment, whereas in the *lumped* models the computational scale is essentially of the same order as that of the catchment.

The main feature of the distributed approach is that the basin outflow is obtained by tracking the water through its different transport phases in the basin interior. In brief, these phases are surface and subsurface transport into the stream channel network, in response to precipitation after it reaches the ground surface, and the subsequent open channel flow to the basin outlet; between precipitation episodes the basin outflow is dominated by baseflow and evaporation processes. The different mechanisms in each of these transport phases may be described by combining some of the formulations of the relevant processes, as presented in Chapters 2–10. These formulations invariably involve a number of assumptions neglecting certain aspects of the flow, which are considered to be less important; this means that they can be only simplified representations of reality. The distributed approach has been receiving increasing acceptance in recent years with the advent of digital computation and with the growing availability of higher-resolution data from digital terrain and other geographical information systems; rapid advances continue to be reported in the literature.

Among the main advantages of distributed models one can note that they allow the exploration of the consequences of various simplifying assumptions; as a result, they can lead to a better understanding of the various pathways and of the interplay between the main processes and related aspects of complex hydrologic systems in the real world. They can also be useful in the prediction of outflow from headwater catchments, provided their parameters can be determined. But this requirement subsumes also one of their main shortcomings. Ideally, the parameters should be determined a priori, that is independently from the model's performance. In many cases, however, this is impossible and the parameters must be estimated by calibration. But then, distributed models tend to contain so many parameters that it becomes practically impossible to estimate them all in objective and physically consistent ways. Another major drawback is that the underlying mathematical rigor of the parameterizations of the model components may instill in the practitioner a confidence and a sense of realism about their performance, that they do not deserve, on account of the many simplifications and uncertainties involved. As a result, the limitations of such models may not be fully understood by uninitiated users and they may be applied to situations for which they were not intended.

In contrast, the lumped models, whose computational scales are of the same order of magnitude as the catchment scales, rely on fewer parameters, which are generally easier to estimate from the available data. Therefore, they are easier to apply in basin outlet flow simulations for prediction and forecasting purposes. Unfortunately, as the computational scale increases, it becomes increasingly difficult to give a physical interpretation to these parameters, in the sense of the processes described in Chapters 2–10. This means that it is usually impossible to predict changes in these parameters, as the catchment undergoes physical changes, such as those resulting from an evolving land use or changing climate. Another drawback is that even when the catchment characteristics remain unchanged, catchment-scale parameters are incapable of accommodating spatial variability of the input (e.g. rainfall) and of the flow processes (e.g. infiltration and evaporation). Moreover, it is impossible to use this approach to describe the detailed flow paths required in the prediction of pollutant transport or erosion. In spite of all these shortcomings, the lumped approach continues to be useful in the prediction of streamflow for certain operational and design purposes. Specific implementations of this approach are further treated in detail in Chapter 12.

Again, in closing this review, it should be understood that, although a classification into distributed and lumped models is useful to bring some order in the multitude of possible approaches, it is also somewhat artificial. Comparison of the different methods treated in Chapters 5–10 has made it clear that the lumped kinematic approach is merely the simplest extreme in a continuous range of complexity levels, which can be applied in up-scaling the analysis from the finest resolution of the full space- and time-dependent conservation equations of momentum, energy and mass to the coarsest resolution, that is the scale of the catchment itself. However, the level of model complexity necessary for a specific application is still not well known; nor is it clear what scenarios warrant the use of more complex models or under what conditions a distributed model will consistently outperform lumped models. In other words, there is still no general consensus regarding the optimal simplifying assumptions that are most appropriate to describe streamflow generation under a given set of conditions. Although it could be argued that there never will be a consensus, this field is in an active state of development and rapid advances continue to be made.

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