Chapter 2
Runoff Generation Mechanisms

CHAPTER 2: RUNOFF GENERATION MECHANISMS

Figure 2 depicts a cross section through a hillslope that exposes in more detail the pathways infiltrated water may follow. Infiltrated water may flow through the matrix of the soil in the inter-granular pores and small structural voids. Infiltrated water may also flow through larger voids referred to as *macropores*. Macropores include pipes that are open passageways in the soil caused by decaying roots and burrowing animals. Macropores also include larger structural voids within the soil matrix that serve as preferential pathways for subsurface flow. The permeability of the soil matrix may differ between soil horizons and this may lead to the build up of a saturated wedge above a soil horizon interface. Water in these saturated wedges may flow laterally through the soil matrix, or enter macropores and be carried rapidly to the stream as subsurface stormflow in the form of interflow.

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Overview of processes involved in runoff generation

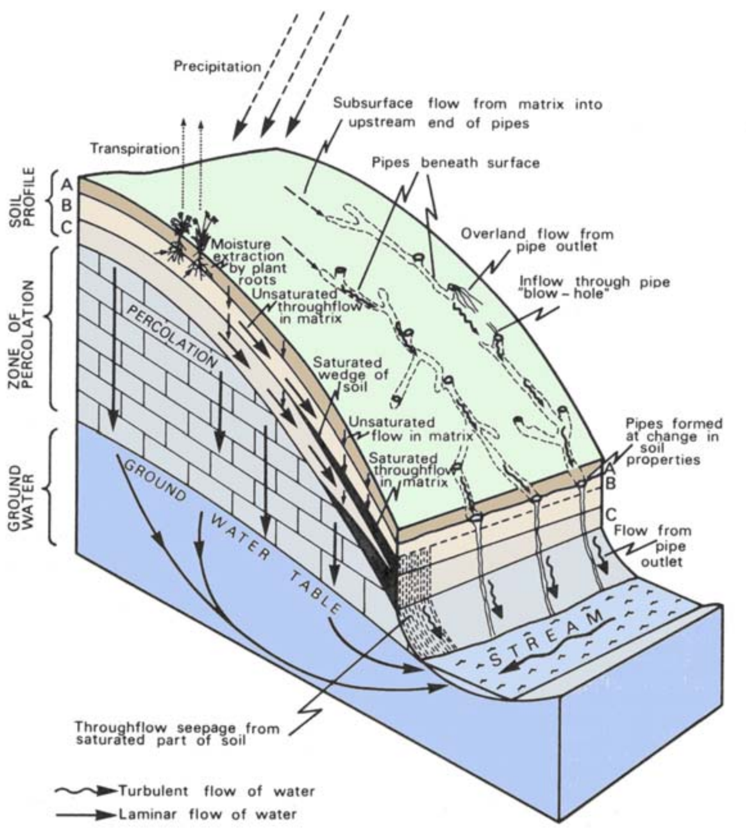

Figure 2. Pathways followed by subsurface runoff on hillslopes. (From Kirkby, 1978)

Recent research in hillslope hydrology involving tracers, especially in humid catchments has found that the dominant contributor to stormflow in the stream is pre-event water (averaging 75% world wide, Buttle, 1994). Pre-event water is water that was present in the hillslope before the storm as identified by a distinct isotopic or chemical composition. Another consensus emerging from recent research is that interflow involving preferential flow through macropores is a ubiquitous phenomenon in natural soils. Rapid lateral flow through a network of macropores and the effusion of old water into stream channels is the primary mechanism for runoff generation in many humid regions where overland flow is rarely observed. This mechanism has been linked to nonlinear threshold type behavior in hillslope runoff response. Figure 3 shows how runoff ratio, the fraction of precipitation that appears as runoff, is dependent upon soil moisture content. Soil moisture content needs to exceed a threshold before any significant runoff occurs. Figure 4 shows the relationship between depth to groundwater and runoff at two different hillslope locations (Seibert et al., 2003) that also shows threshold behavior, with runoff being more tightly related to depth to groundwater near the stream than further up a hillslope.

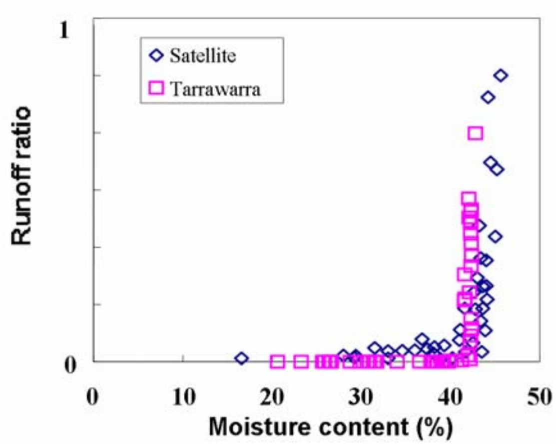

Figure 3. Relationship between runoff ratio and soil moisture content (Woods et al., 2001, Copyright, 2001, American Geophysical Union, reproduced by permission of American Geophysical Union).

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Animation of Preferential Pathway Infiltration

Natural soils contain heterogeneities that lead to variability in the infiltration process itself. Infiltrating water follows preferential pathways and macropores and may result in increases in moisture content at depth before saturation or similar increases in moisture content higher in the soil profile. Figure 5a shows a photograph of a soil where dye has been used to trace infiltration pathways in experiments reported by Weiler and Naef (2003). Figure 5b shows the dye intensity objectively classified from the photograph following excavation of the plot following a dye sprinkling experiment. Figure 5c shows moisture content over time measured at a range of depths using time domain reflectometry in these sprinkler experiments.

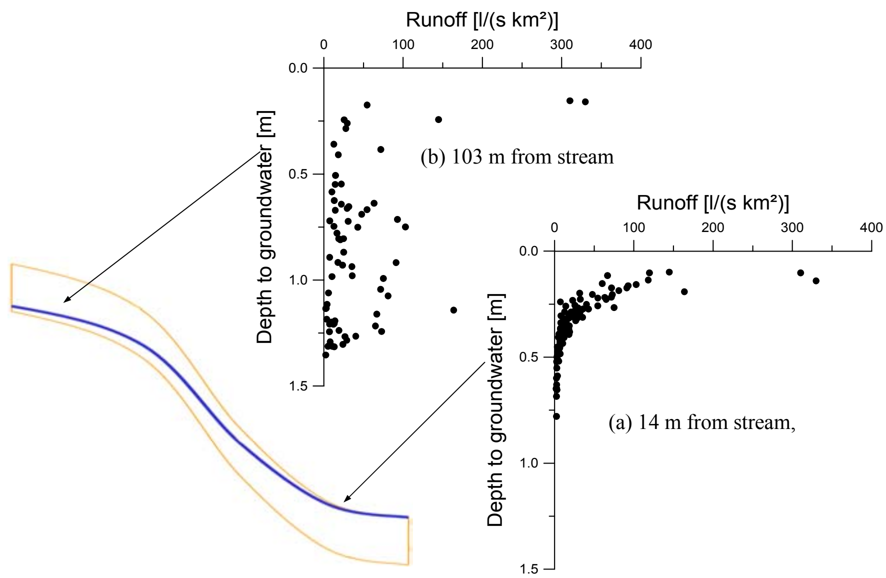

Figure 4. Relation between runoff and depth to groundwater for two different locations in the Svartberget catchment (Seibert et al., 2003, Copyright, 2003, American Geophysical Union, reproduced by permission of American Geophysical Union)

With this background on the pathways followed by infiltrated water we can examine the mechanisms involved in the generation of runoff (Figure 6). Each mechanism has a different response to rainfall or snowmelt in the volume of runoff produced, the peak discharge rate, and the timing of contributions to streamflow in the channel. The relative importance of each process is affected by climate, geology, topography, soil characteristics, vegetation and land use. The dominant process may vary between large and small storms.

Figure 5. (a) Photograph of cross section through soil following dye tracing experiment (Courtesy of Markus Weiler).

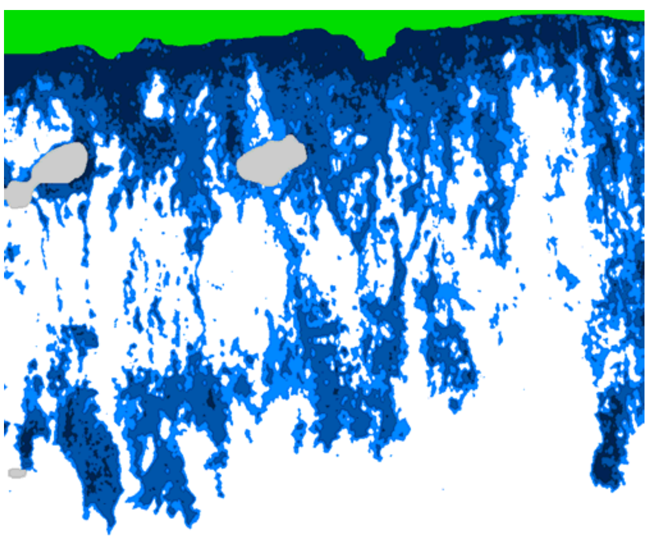

Figure 5. (b) Objectively classified dye intensity following sprinkler experiment. (Courtesy of Markus Weiler)

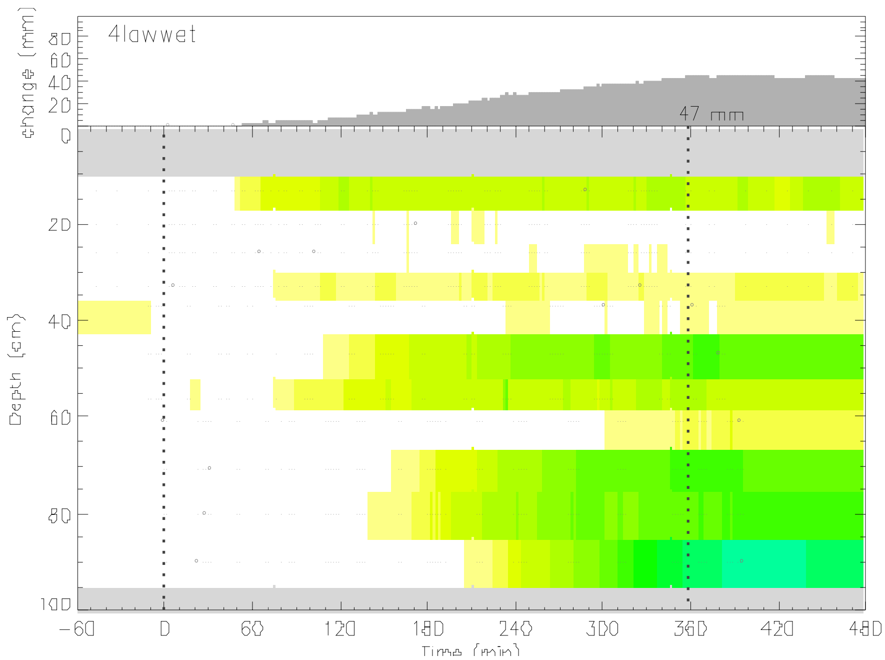

Figure 5. (c) Moisture content change measured using time domain reflectometry during sprinkler experiment. (Courtesy of Markus Weiler)

In Figure 6a the *infiltration excess* overland flow mechanism is illustrated. There is a maximum limiting rate at which a soil in a given condition can absorb surface water input. This was referred to by Robert E. Horton (1933), one of the founding fathers of quantitative hydrology, as the *infiltration capacity* of the soil, and hence this mechanism is also called Horton overland flow. Infiltration capacity is also referred to as *infiltrability*. When surface water input exceeds infiltration capacity the excess water accumulates on the soil surface and fills small depressions. Water in depression storage does not directly contribute to overland flow runoff; it either evaporates or infiltrates later. With continued surface water input, the depression storage capacity is filled, and water spills over to run down slope as an irregular sheet or to converge into rivulets of overland flow. The amount of water stored on the hillside in the process of flowing down slope is called *surface detention*. The transition from depression storage to surface detention and overland flow is not sharp, because some depressions may fill and contribute to overland flow before others. Figure 7 illustrates the response, in terms of runoff from a hillside plot due to rainfall rate exceeding infiltration capacity with the filling of depression storage and increase in, and draining of, water in

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Animations of Infiltration Excess Runoff Generation surface detention during a storm. Note, in Figure 7, that infiltration capacity declines during the storm, due to the pores being filled with water reducing the capillary forces drawing water into pores.

Due to spatial variability of the soil properties affecting infiltration capacity and due to spatial variability of surface water inputs, infiltration excess runoff does not necessarily occur over a whole drainage basin during a storm or surface water input event. Betson (1964) pointed out that the area contributing to infiltration excess runoff may only be a small portion of the watershed. This idea has become known as the *partial-area* concept of infiltration excess overland flow and is illustrated in Figure 6b.

Infiltration excess overland flow occurs anywhere that surface water input exceeds the infiltration capacity of the surface. This occurs most frequently in areas devoid of vegetation or possessing only a thin cover. Semi-arid rangelands and cultivated fields in regions with high rainfall intensity are places where this process can be observed. It can also be seen where the soil has been compacted or topsoil removed. Infiltration excess overland flow is particularly obvious on paved urban areas.

In most humid regions infiltration capacities are high because vegetation protects the soil from rain-packing and dispersal, and because the supply of humus and the activity of micro fauna create an open soil structure. Under such conditions surface water input intensities generally do not exceed infiltration capacities and infiltration excess runoff is rare. Overland flow can occur due to surface water input on areas that are already saturated. This is referred to as *saturation excess* overland flow, illustrated in Figure 6c. Saturation excess overland flow occurs in locations where infiltrating water completely saturates the soil profile until there is no space for any further water to infiltrate. The complete saturation of a soil profile resulting in the water table rising to the surface is referred to as *saturation from below*. Once saturation from below occurs at a location all further surface water input at that location becomes overland flow runoff.

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Animation of Saturation Excess Runoff Generation

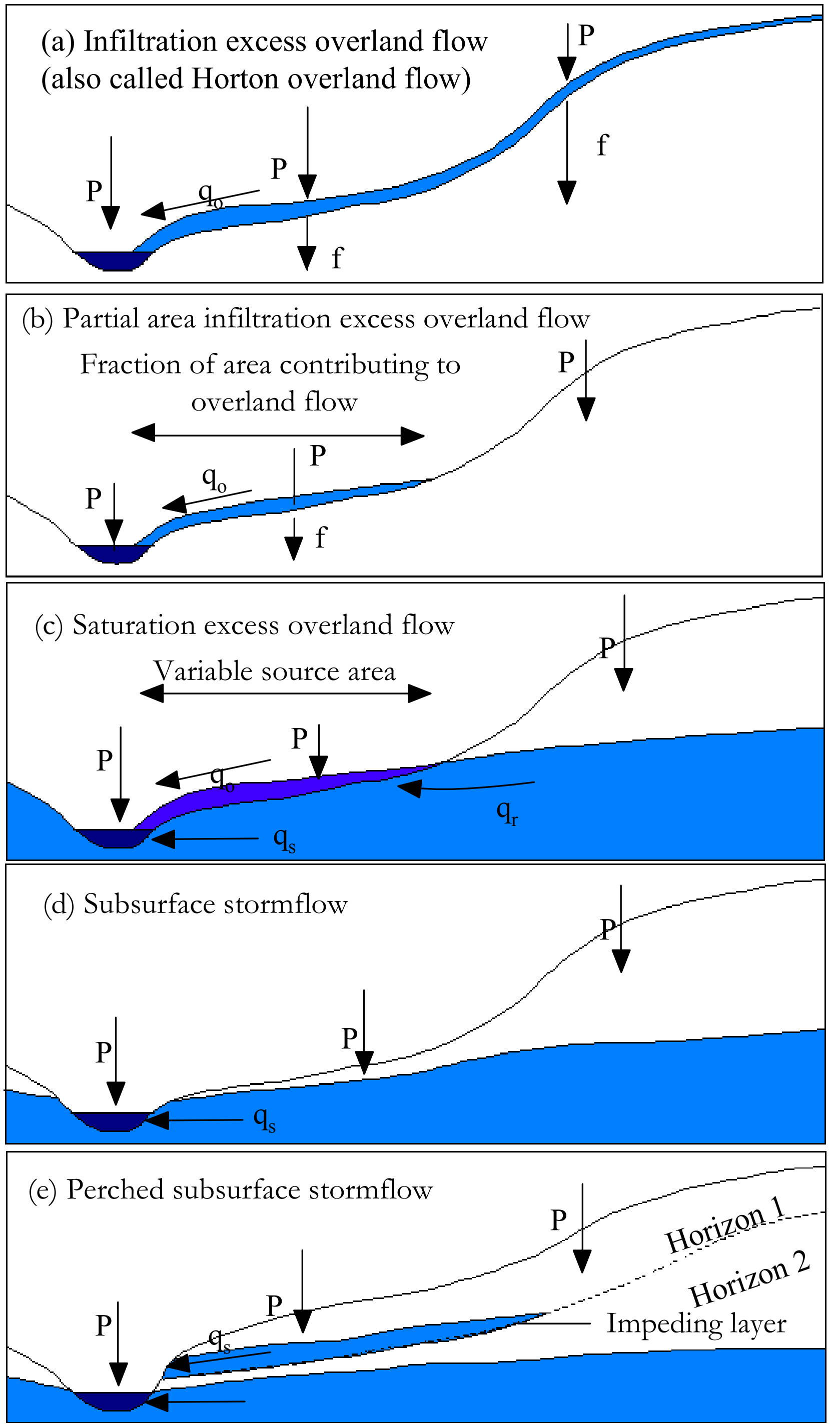

Figure 6. Classification of runoff generation mechanisms (following Beven, 2000)

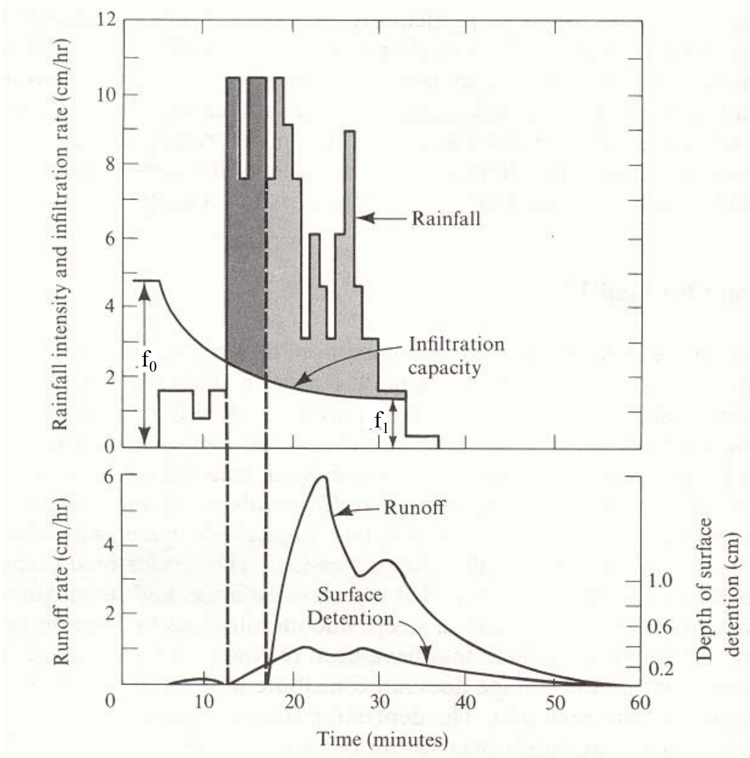

Figure 7. Rainfall, runoff, infiltration and surface storage during a natural rainstorm. The shaded areas under the rainfall graph represent precipitation falling at a rate exceeding the infiltration rate. The dark grey area represents rainfall that enters depression storage, which is filled before runoff occurs. The light grey shading represents rainfall that becomes overland flow. The initial infiltration rate is f_0, and f_1 is the final constant rate of infiltration approached in large storms (from Water in Environmental Planning, Dunne and Leopold, 1978).

In humid areas streams are typically gaining streams (gaining water by drainage of baseflow from the groundwater into the stream) with the groundwater table near the surface coincident or close to the stream water surface elevation. This means that the water table near streams is close to the ground surface, especially in flat topography, making these near stream areas in flat topography particularly susceptible to saturation from below. The extent of the area subject to saturation from below varies in time, both at seasonal and event time scales due to fluctuations in the depth to the shallow water table. This variability of the extent of surface saturation is referred to as the *variable source area* concept (Hewlett and Hibbert, 1967) and is illustrated in Figures 8 and 9.

Geometrical considerations dictate that near stream saturated zones will be most extensive in locations with concave hillslope profiles and wide flat valleys. However, saturated overland flow is not restricted to near-stream areas. Saturation from below can also occur (1) where subsurface flow lines converge in slope concavities (hillslope hollows) and water arrives faster than it can be transmitted down slope as subsurface flow; (2) at concave slope breaks where the hydraulic gradient inducing subsurface flow from upslope is greater than that inducing down slope transmission; (3) where soil layers conducting subsurface flow are locally thin; and (4) where hydraulic conductivity decreases abruptly or gradually with depth and percolating water accumulates above the low-conductivity layers to form perched zones of saturation that reach the surface.

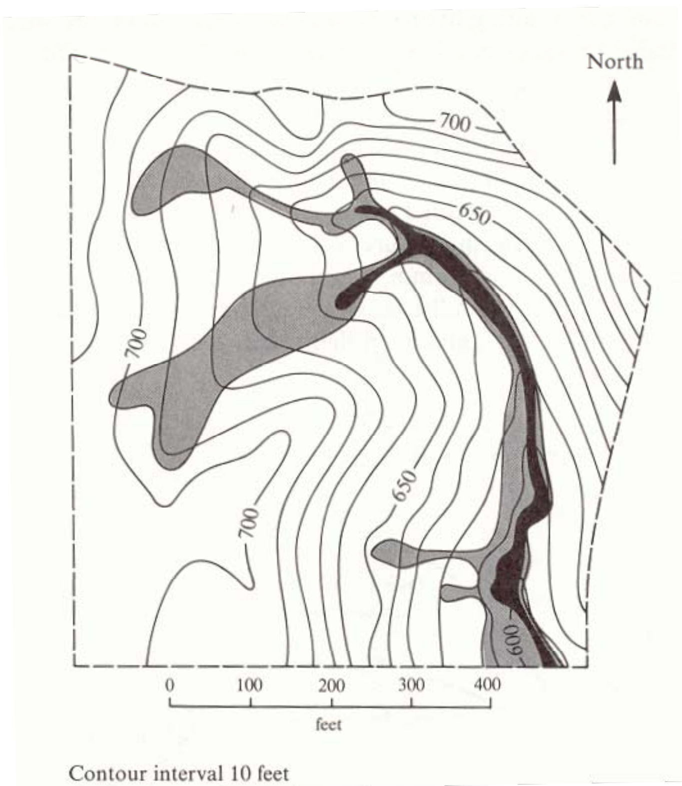

Figure 8. Map of saturated areas showing expansion during a single rainstorm. The solid black shows the saturated area at the beginning of the rain; the lightly shaded area is saturated by the end of the storm and is the area over which the water table had risen to the ground surface (from Water in Environmental Planning, Dunne and Leopold, 1978)

Return flow (q_r in Figure 6c) is subsurface water that returns to the surface to add to overland flow. Return flow also occurs at places where the soil thins, for example rock outcrops and may manifest in the form of springs.

In areas with high infiltration capacities, interflow, or subsurface storm flow is usually the dominant contributor to streamflow, especially on steeper terrain or more planar hillslopes where saturation excess is less likely to occur. A number of processes are involved in rapid subsurface stormflow. These include *transmissivity feedback*, *lateral flow at the soil bedrock interface* and *groundwater ridging*.

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Animation of Subsurface Stormflow

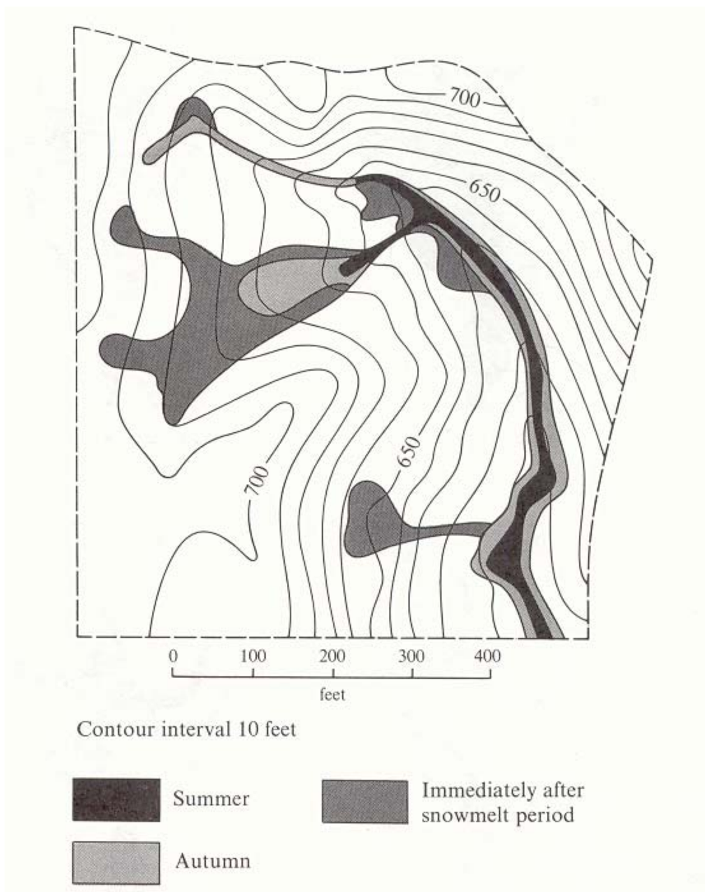

Figure 9. Seasonal variation in pre-storm saturated area (from Water in Environmental Planning, Dunne and Leopold, 1978)

Transmissivity feedback (Weiler and McDonnell, 2003) is illustrated in Figure 10 and occurs when water infiltrates rapidly along preferential pathways and causes the groundwater to rise to the point where highly permeable soil layers or macropore networks become activated and transmit water rapidly downslope. Much of the water that drains from the soil matrix into the macropore network is pre-event water. This mechanism results in a nonlinear threshold like response as illustrated in Figures 3 and 4.

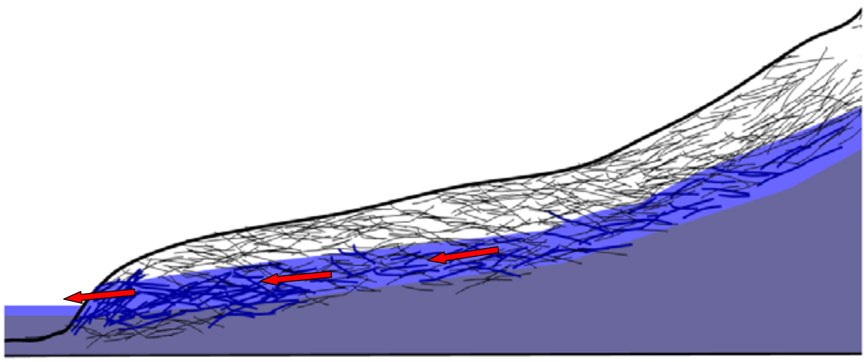

Figure 10. Schematic illustration of macropore network being activated due to rise in groundwater resulting in rapid lateral flow.

Lateral flow at the soil bedrock interface (Weiler and McDonnell, 2003) illustrated in Figure 11, occurs in steep terrain with relatively thin soil cover and low permeability bedrock, where water moves to depth rapidly along preferential infiltration pathways and perches at the soil-bedrock interface. Since moisture content near the bedrock interface is often close to saturated, the addition of only a small amount of new water (rainfall or snowmelt) is required to produce saturation at the soil-bedrock or soil-impeding layer interface. Rapid lateral flow occurs at the permeability interface through the transient saturated zone. Once rainfall inputs cease, there is a rapid dissipation of positive pore water pressures and the system reverts back to a slow drainage of matrix flow.

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Animation of Perched layer stormflow

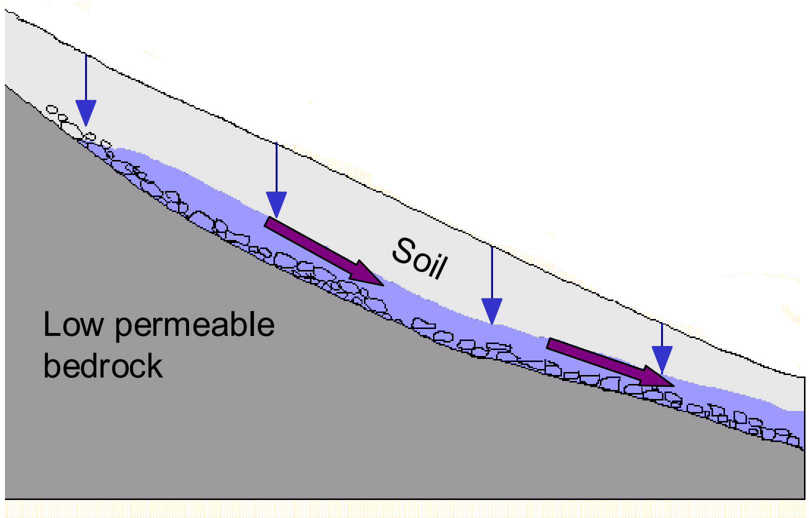

Figure 11. Rapid lateral flow at soil bedrock interface.

The processes involved in the generation of subsurface stormflow by groundwater ridging are illustrated in Figure 12. An idealized cross section of a valley with a straight hillslope is shown. In a simplified situation with uniform soils the water table has an approximately parabolic form, and soil moisture content decreases with increasing height above the water table. The shaded areas represent graphs of soil moisture at the base, middle and near the top of the hillslope (a) before the onset of rainfall; (b) as an initial response to rainfall; and (c) after continuing rainfall. Because (in a) before the onset of water input the water table slopes gently towards the channel there will be a slow flow of groundwater to maintain the baseflow of the stream. With the onset of surface water input, water that infiltrates near the base of the hillslope will quickly reach the water table and cause the water table near the stream to rise, early in a storm. Further upslope the soil is dryer and distance to the water table greater. It therefore takes longer for infiltrating water to reach the water table and where the water table is deep all the infiltrating water may go into storage in the unsaturated zone and not reach the water table for many days after the storm. The initial response to water input is therefore as depicted in Figure 12b, where the water table has risen near the stream but remained unchanged further upslope. The rising water table near the stream causes an increase in the hydraulic gradient between the groundwater and stream, and increased subsurface flow into the stream results. This is subsurface stormflow, and is frequently seen to be groundwater that has been displaced by the infiltrating water, and is thus old or pre-storm water bearing the chemical and isotopic signature of water in the hillslope prior to the storm, which may be different from the chemical and isotopic signature of overland flow from rainwater that has not infiltrated. Measurement of chemical and isotopic signatures of stream water, ground water and rain water is commonly used in hydrology as a way of inferring hillslope flow pathways. After continuing rain (Figure 12c), the water table has risen to the surface over the lower part of the hillslope and the saturated area is expanding uphill. Some water emerges from this saturated area and runs down slope to the stream. This is termed *return flow*. Direct precipitation onto the saturated zone (DPS) forms saturation excess runoff as described above.

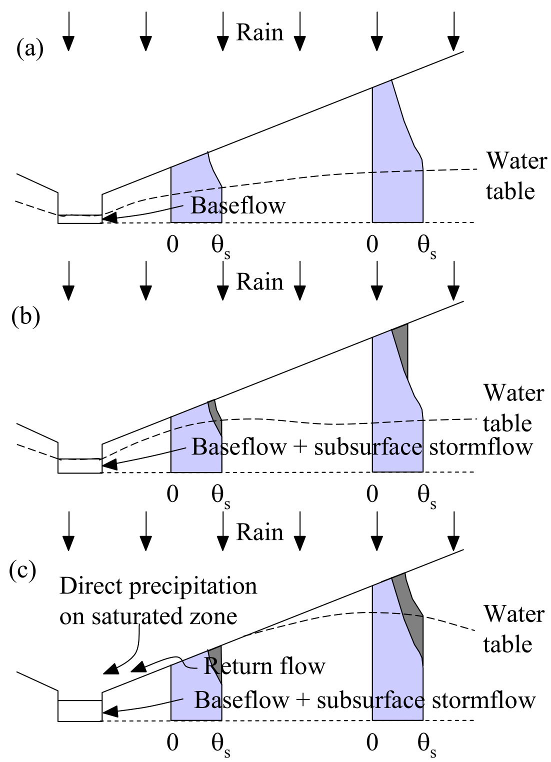

Figure 12. Groundwater ridging subsurface stormflow processes in an area of high infiltration rate. (redrawn following Water in Environmental Planning, Dunne and Leopold, 1978)

Figure 12 illustrates a region just above the water table that was close to saturation. This is known as the *capillary fringe*, and can play an important role in runoff generation in certain situations. Capillary forces due to the surface tension between water and soil particles act to pull water into the soil matrix above the water table and maintain the capillary fringe at moisture content very close to saturation. The addition of a small amount of water can saturate this soil and cause the water table to rise quite rapidly, resulting in subsurface stormflow, surface saturation and saturation excess overland flow. The moisture content in the capillary fringe can also be affected by the history of wetting and drying of the soil, a phenomenon known as *hysteresis*. When soil has been draining the moisture content tends to remain above what it would be if it were filling at the same pressure. The addition of a small amount of water can switch the soil from draining to filling mode, enhancing the effect of the capillary fringe on the rise of the water table and subsurface stormflow response. The capillary fringe and hysteresis are discussed in more detail in Chapter 4.

The discussion thus far has focused on the main processes involved in runoff generation on a hillslope. To complete the discussion on runoff generation processes it is necessary to mention briefly some other processes and factors involved. Interception of precipitation by vegetation can play a significant role in reducing runoff, especially in forested environments. Much intercepted water is eventually evaporated back to the atmosphere (Figure 1). In some hydrologic models, interception is sometimes modeled as an *initial abstraction* that is subtracted from precipitation inputs before they are used in infiltration or runoff calculations. In other hydrologic models detailed representations of the interception, storage of water in the canopy, throughfall or stem flow are used (e.g. Rutter et al., 1972).

Direct precipitation onto a stream or water body also contributes to runoff as indicated in Figure 6. This is important in areas where the water surface is extensive, as with lakes, reservoirs and floodplains that are flooded, because in these situations runoff generation is not delayed by the usual hillslope processes.

The freezing state of the soil, in regions where freezing occurs, also plays a role in runoff generation. Infiltration capacity is reduced due to frozen ground, depending upon the soil moisture content at the time of freezing.

Fire results in water repellency by soils which reduces infiltration capacity. One cause for water repellency is chemicals released during a fire that are absorbed in the soil, and can make it water repellent for months to years following a fire. The heat from fire also removes the thin films of irreducable water adhered to soil particles by capillary forces, disconnecting potential flow paths. Penetration of water into macropores following a fire is limited due to this effect. High temperatures in deserts have the same effect, adding to the tendency for infiltration capacities to be lower in arid regions making them more subject to infiltration excess runoff generation processes. This water repellency due to fire has been implicated in many floods following severe bush or forest fires.

Many of the runoff generation processes described depend on the soil moisture status of the soil. This is referred to as the *antecedent conditions*. Between storms (surface water input events), processes of evaporation, transpiration, percolation and drainage serve to set up the soil moisture antecedent conditions. Runoff generation mechanisms and processes therefore depend not only on conditions during storms, but conditions in advance of storms and a complete understanding or representation of all the land surface hydrologic processes is required to quantify the generation of runoff. Recognition of this has led to the development of continuous simulation models, such as the National Weather Service Sacramento soil moisture accounting model that keeps continuous track of the state of different soil moisture components for the modeling of runoff. Detailed presentation of these models is beyond the scope of this module, although key ideas are reviewed at the end of this model.

The discussion above has reviewed, in a conceptual way many of the processes and mechanisms involved in runoff generation. These can be quite complex, and when efforts are made to perform quantitative calculations the devil is in the details. Each watershed or hillslope is different, with different topography, soils and physical properties. The challenge for hydrologic modelers is to balance practical simplifications with justifiable model complexity and the knowledge that many specific physical properties required for detailed hydrologic modeling are physically unknowable. Our understanding of runoff generation involves the movement of water through soil pores and macropores. These flows follow the physical laws governing fluid flow (Navier Stokes equations) but we can never know in sufficient detail the flow geometry to make use of fluid flow theory and ultimately have to resort to simplifications or parameterizations of the runoff generation processes. In the remainder of this module the astute reader will note discrepancies between the physical understanding given above and mathematical descriptions used to perform practical calculations. The mathematical descriptions, although frequently complex, incorporate significant simplifications relative to the field based conceptual understanding of how runoff processes work. This gap between field based and model based representations makes the subject of rainfall – runoff processes a fertile area for research to learn how to better model rainfall runoff processes.

Figure 13 summarizes the main processes involved in runoff generation, showing the interaction between infiltration excess, saturation excess and groundwater flow pathways. Most rainfall runoff models are organized around a representation similar to Figure 13 involving partition of surface water input into infiltration or overland flow, either due to infiltration excess or saturation excess. Infiltrated water enters the soil regolith where it contributes to interflow, percolates to deeper groundwater or is evaporated or transpired back to the atmosphere. The quantity of water in the soil affects the variable source area involved in the generation of saturation overland flow. The deeper groundwater contributes to baseflow and affects interflow through groundwater rise.

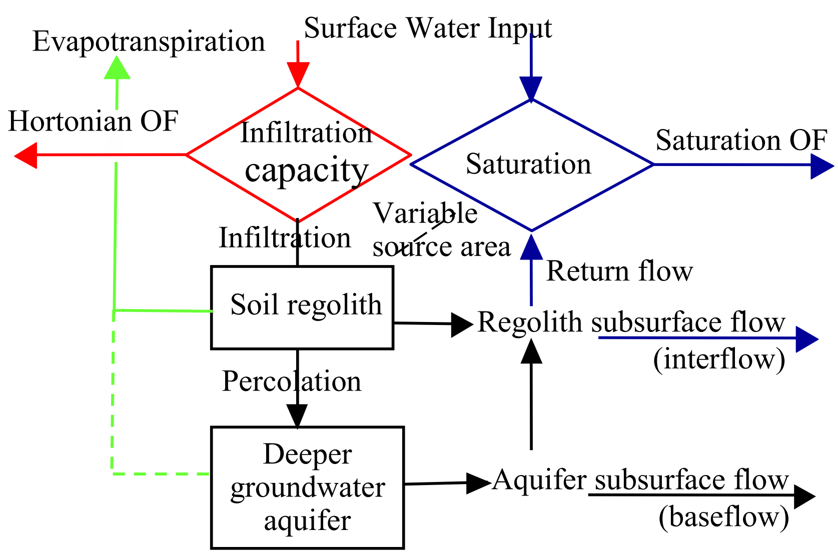

Figure 13. Hydrological Pathways involved in different runoff generation processes. Infiltration excess pathways are shown in red. Saturation excess and subsurface stormflow pathways are shown in blue. Groundwater and baseflow pathways in black and Evapotranspiration is green. (Courtesy of Mike Kirkby)

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Exercises

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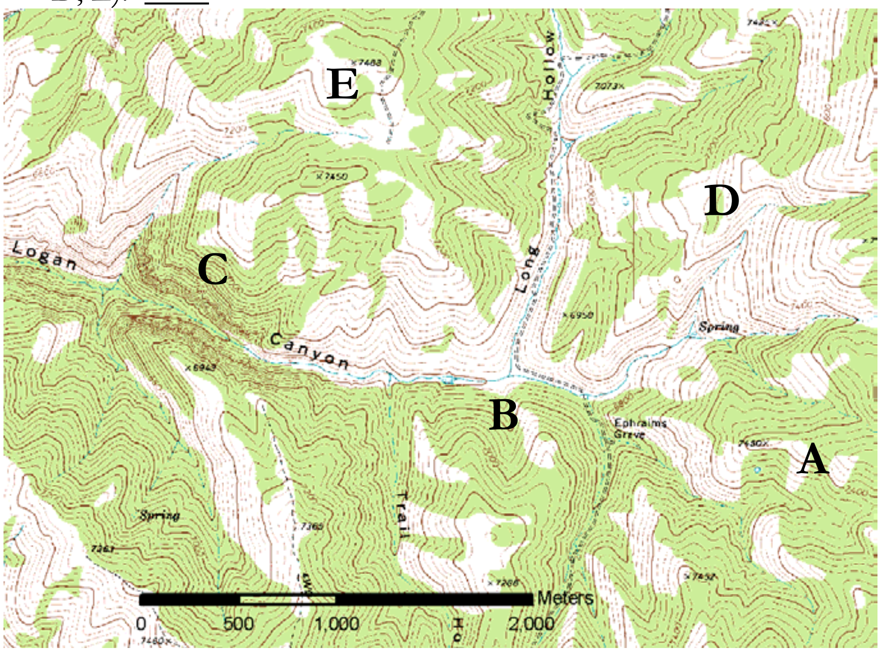
Do the Chapter 2 Quiz

1. Given the topographic map from Logan Canyon below, indicate the location where saturation excess overland flow is most likely to be generated during rainfall (from labeled locations, A, B, C, D, E): ____



2. Infiltration capacity is:
   A. The number of foreign spies that a country can tolerate
   B. The rate of water input to a stream by subsurface flow
   C. The fraction of watershed area contributing to overland flow
   D. The maximum rate at which water can be absorbed into soil
   E. The water holding capacity of surface depressions

3. Subsurface stormflow is likely to be larger in:
   A. A steep narrow valley
   B. A wide flat valley

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