River Basin Hydrology:
Basin surface run-off, measurement of river discharge

Basin Hydrology: flow generation

Inputs of water to a drainage basin are the various forms of **precipitation** that fall over its area. These include rain, snow, sleet, hail and dew. If you look at a drainage basin on a map, you will see that river channels cover only a very small part of the total area. This means that most of the water reaching the ground surface must find its way from the hillslopes and into the channel network. A number of pathways are involved (Figure 1), and a given 'parcel' of water arriving at a stream channel may have taken any number of them. Not all the incoming precipitation actually makes it to the basin outlet, since a certain percentage is evaporated back to the atmosphere. Water that falls on the leaves of vegetation and artificial structures like buildings is **intercepted**. Some of this water does fall to the ground, as anyone who has sheltered under a tree will know, but much of it is evaporated (water resource managers refer to this output as 'interception loss'). The amount of interception that occurs is dependent on such factors as the extent and type of vegetation (leaf size, structure, density and the arrangement of foliage), wind speed and rainfall intensity (Jones, 1997). Evaporation also takes place from the surface layers of the soil, and from lakes and wetland areas. A related process is transpiration, whereby plants take up water through their roots and evaporate it through pores, called stomata, on the underside of their leaves. This means that water can be 'lost' to the atmosphere from some depth below the surface. Because evaporation and transpiration are difficult to monitor separately, they are usually considered together as the combined process of evapotranspiration. Water reaching the soil surface may either enter the soil by a process called infiltration or remain at the surface, moving down-slope as overland flow (labeled 1 and 4 in Figure 1). Infiltrated water either travels through the soil, parallel to the surface, as throughflow (2), or slowly percolates downwards to the saturated zone, travelling as groundwater flow (3). Rates of water movement are fastest for overland flow (typically between 50 m and 500 m per hour), between 0.005 m and 0.3 m per hour for throughflow and 0.005 m to 1.5 m per day for groundwater flow (Ward and Robinson, 1990).

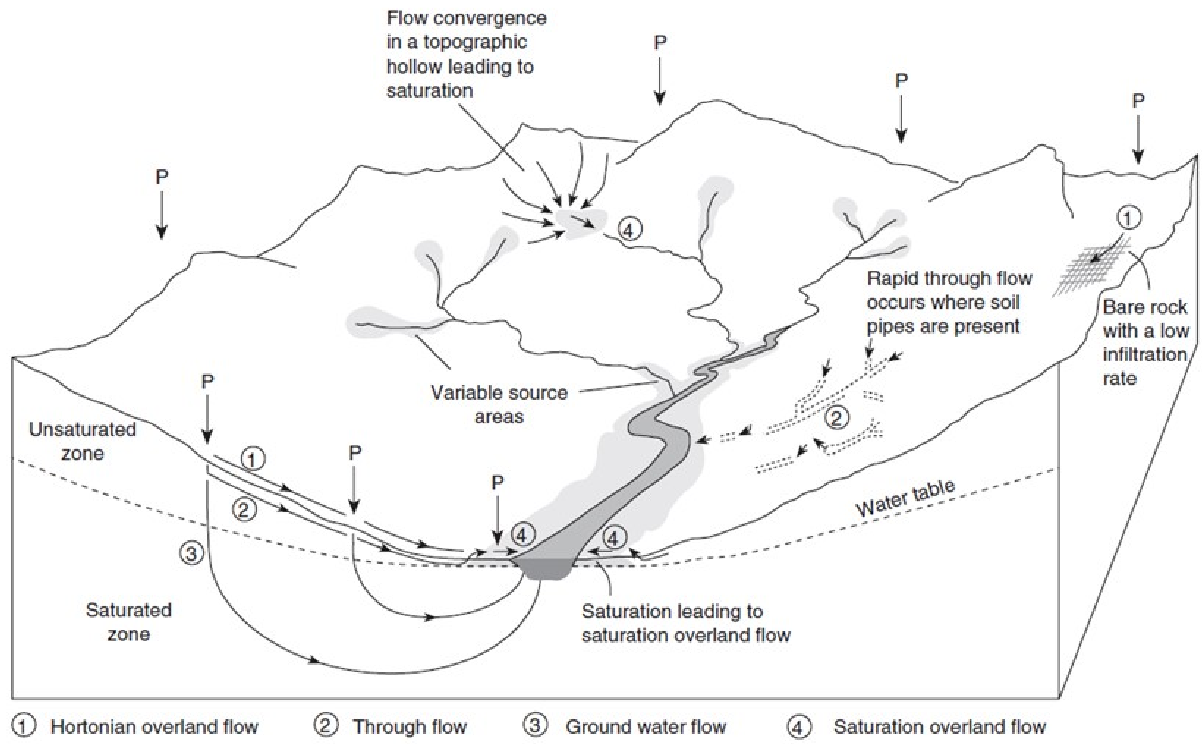

Figure 1 Surface and subsurface hydrological pathways.

How fast a river rises after rainfall is very much dependent on the relative proportion of water taking faster (surface and near-surface) pathways, and slower (subsurface) pathways. An important control on this is the rate at which water infiltrates into the soil, which is determined by the infiltration capacity. This is defined as the maximum rate at which water enters the soil when it is in a given condition (Horton, 1933). This is an important definition: if the rainfall intensity exceeds the infiltration capacity, the soil cannot absorb all the water and there is excess water, which ponds at the surface. This moves down slope as overland flow, more specifically, Hortonian overland flow (1). As Horton's definition implies, soil properties control the infiltration capacity; these include soil permeability, the presence of vegetation and plant roots and how much water is already in the soil. Following a dry period, the infiltration capacity is highest at the start of a storm, rapidly decreasing as rainfall continues, until a constant infiltration rate is reached (Horton, 1933).

A second type of overland flow, saturation overland flow (4), is generated when the soil is totally saturated and therefore cannot take any more water (Kirkby and Chorley, 1967). Where the water table is relatively shallow, for example in valley bottoms, rainfall can cause it to rise to the ground surface. Where this occurs, a saturated area forms around the channel, increasing in extent as the storm progresses. The saturated area acts as an extension to the channel network, meaning that a significant volume of water is transferred in a short time. These saturated areas are known as variable source areas or dynamic contributing areas and are highly significant in humid environments, where this is the main way in which storm run-off is generated (Hewlett and Hibbert, 1967; Dunne and Black, 1970). Hortonian overland flow is rarely observed in humid environments, unless the surface has a low infiltration rate, for example where there are outcrops of bare rock or artificially paved surfaces. However, in dryland environments, a combination of factors mean that Hortonian overland flow is the dominant mechanism. Rainfall, when it occurs, typically has a high intensity and exceeds the low infiltration capacity of sparsely vegetated soils (Dunne, 1978). In addition, dryland soils often develop a crust at the surface, which further reduces the infiltration capacity.

Water that infiltrates the soil may travel at various depths below the surface. While it is fairly obvious that water should move downwards under the force of gravity, it might seem counter-intuitive that it also flows through the soil as throughflow. This happens because a preferential flow path is set up. Soil permeability decreases with depth, meaning that the downward movement of water is slowed. During rainfall, this leads to a backing up effect in the more permeable surface layers. This has been likened to the flow of water down a thatched roof: it is easier for the water to move parallel to the slope of the roof, along the stems of the straw, than to move vertically downwards through it (Ward, 1984; Zaslavsky and Sinai, 1981).

Where they exist, soil pipes provide a very rapid throughflow mechanism, and rates of flow can be comparable to those in surface channels. Soil pipes are hydraulically formed conduits that can be up to a metre or more in diameter. They are found in a wide range of environments and are sometimes several hundreds of metres in length (Jones, 1997). As such, they can act as an extension to the channel network, allowing the drainage basin to respond rapidly to precipitation inputs (Jones, 1979).

The schematic diagram in Figure 1 represents the headwaters of a humid zone river, where the channel intersects the water table surface (it should be noted that aquifers are not always present, for example aquifer development is extremely limited in headwater areas that are underlain by impermeable rocks and characterised by thin soils). In this example, groundwater contributions are made to the channel flow from the underlying aquifer and the channel is called a gaining stream. Even during rainless periods, flow will be maintained, as long as the level of the water table does not fall below that of the channel bed. A rather different situation exists for many dryland channels, where the water table may be several metres, or tens of metres, below the surface. In this case, the direction of flow is reversed, as water is lost through the bed and banks of the channel, percolating downwards to recharge the aquifer. This is termed a losing stream and, although not exclusive to drylands, many examples are found in these environments. It is not unusual for a river to be gaining and losing flow to groundwater along different parts of its course, while seasonal fluctuations in water table levels can mean that losing streams become gaining streams for part of the year. The Euphrates in Iraq provides a good example, with most of the flow being generated in the headwaters in northern Iraq, Turkey and Syria. Further downstream, at the Hit gauging station (150 km west of Baghdad) the river loses flow to groundwater for much of the year (Wilson, 1990). The loss of flow from a channel, due to downward percolation and high evaporation rates, is referred to as transmission loss.

Reference: Charlton Ro (2007). FUNDAMENTALS OF FLUVIAL GEOMORPHOLOGY