Groundwater principles

Origin of groundwater

All earth's water was formed deep underground by magmatic processes, and has over aeons been released at the surface and on ocean floors by volcanism. The mechanism continues today. With the exception of this 'new' water, all groundwater is derived from that part of precipitation which, after surface runoff and evaporation, infiltrates the soil. Some of the infiltrating water is transpired by plants, some is drawn upward by capillary action and evaporated, and some remains indefinitely in microscopic voids in the soil profile. During and after continuous and wetting rain, the remainder infiltrates downwards, intermittently and successively saturating the material through which it passes, until the water reaches the zone of saturation. Here, the soil or rock voids (openings) are completely filled with water. The water is then called groundwater, and the upper surface of the zone of saturation is known as the water table. The water table is usually a subdued replica of the land surface, being almost flat under gently undulating ground and deeper and sloping under hills.

The proportion of rain infiltrating into the soil is very variable, ranging from a few percent on steep, rocky slopes, to perhaps 50% or more in sandy or gravelly areas with little runoff. The proportion also changes seasonally, and infiltration would be expected to be a maximum when evaporation is least – at night in winter. Of the water which enters the soil, only a fraction avoids transpiration or retention in soil voids, and infiltrates to the water table.

Groundwater is therefore a part of the general hydrological cycle, and is directly related to the surface movement of water.

Unconfined and confined aquifers

An aquifer is a body of rock, or unconsolidated material such as sand, capable of supplying useful amounts of groundwater. An aquifer has two purposes: it stores, and transmits, groundwater. The relative importance of each function is determined by the nature of each aquifer. Some aquifers (eg hard sandstone) may store only a small amount of water in a network of thin fractures, but might transmit it freely, and remain reliable suppliers, if the fractures are sufficiently interconnected. Other materials like fine-grained and porous clays may contain larger amounts of water, but yield only small amounts because the water is not transmitted easily through their microscopic voids.

Aquifers may be unconfined, confined or semi-confined. An unconfined or water table aquifer exists in unconsolidated sediments or hard, fractured rock whenever the water table is in contact with air at atmospheric pressure. Unconfined aquifers therefore receive recharge from infiltrating rain over their full areal extent. Groundwater in a bore tapping an unconfined aquifer is encountered at the level of the water table. A bore drilled into an unconfined fractured rock aquifer may remain dry to depths below the water table if no water-bearing fractures are intersected[1], but once they are, the water will rise to the level of the water table. Since fractured rock aquifers are largely solid, dry rock separated by a network of fractures, it is possible for two bores side by side to yield different amounts of water, or either or both might remain dry.

A confined aquifer is a saturated, permeable zone bounded above and below by relatively impermeable materials (rock or soil). The zone therefore cannot receive recharge by directly infiltrating rain, but must get it from a recharge area elsewhere, where the permeable zone is exposed at the land surface, and where at least local unconfined conditions exist. The infiltrating groundwater in the zone of recharge moves crossgradient or downgradient beneath the confining impermeable layer. The water in confined zones of aquifers is therefore not in contact with the atmosphere, and is at a pressure greater than atmospheric. Water in bores tapping confined aquifers rises up the bore under pressure, and may overflow at the land surface. If the water in the bore rises above the land surface (so that groundwater flows without the need for a pump), the groundwater (and the bore) are said to be artesian. If the groundwater rises but not sufficiently for the bore to flow, the groundwater is sub-artesian.

[1] At this local scale, groundwater conditions are confined.

A semi-confined aquifer receives vertical groundwater leakage from a higher aquifer down via a semi-permeable (rather than impermeable zone) zone separating them.

It is possible for an aquifer to be unconfined in one part of it, confined in another, and semi-confined elsewhere. The zone of confinement or semi-confinement may be relatively small, so that locally the aquifer behaves in a confined manner, but on a broader scale, unconfined conditions dominate. An example is a fractured hard rock aquifer where water is contained only within joints and similar defects which extend and are open to the land surface, separated by impermeable rock where no water is present. The water in the joints is unconfined. Drilling through the rock produces no water, which is only struck (and which rises to the level of the water table) when a water bearing fracture is intersected.

Storage capabilities of fractured rock aquifers

Groundwater in fractured rock aquifers is stored in fractures within the rock mass. Usually, the volume of fractures as a proportion of the rock mass is low, and commonly less than a few percent.

These aquifers therefore often have low storage capabilities, in comparison to unconsolidated aquifers like coastal sands. In these materials, the water is stored in voids between the sand grains, and the voids are interconnected (ie the aquifer is intergranular). The voids may constitute from 25% to 35% of the volume of sand (ie the porosity, θ, of the sand is 25% to 35%, or 0.25 to 0.35 expressed as a fraction). Each cubic metre of saturated sand below the water table therefore contains 250L to 350L of groundwater.

Primary and secondary porosity

The voids between sand grains in a coastal sand body, or the vesicles in otherwise hard basalt, for example, constitute primary porosity, because they were formed at the same time as the sand was deposited, or the basalt flowed as lava. As the sand becomes progressively cemented and consolidated in the process of becoming hard rock, the primary porosity is reduced. Most hard rocks have very little remaining primary porosity. However, if the hard rock becomes fractured and otherwise jointed, the fractures constitute secondary porosity.

Groundwater gradient

Groundwater is rarely stationary. It moves in response to gravity, and hydrostatic and lithostatic pressures, from recharge areas to discharge zones. Discharge occurs wherever the water table intersects the land surface in springs, swamps, rivers and the sea, provided the water table slopes towards the feature. If the water table is lower than the feature, water may flow from the spring or river to the groundwater body. The slope of the water table is called the water table gradient[2], which determines the direction and rate at which groundwater moves. The greater the gradient, the more rapid the flow. Groundwater usually flows in the direction of steepest gradient.

Aquifer hydraulic conductivity and transmissivity

Hydraulic conductivity (symbol K) is a measure of how readily an aquifer transmits water, and is defined as the rate at which groundwater will flow from a unit area (eg one square metre) of aquifer under a unit gradient (ie the gradient is 1). It is expressed as cubic metres per day per square metre (m^3/day/m^2, which reduces to m/day).

Permeabilities of fractured rock aquifers are a function of the intensity of fracturing, their openness, and the degree to which they interconnect. Since these features are often very variable, hydraulic conductivity also varies widely. Typical ranges for fractured, hard rock might be 0.01 – 100m/day. Transmissivity (T) is defined as the product of hydraulic conductivity and saturated aquifer thickness, and is therefore the rate at which groundwater will flow from a vertical, one-metre wide strip of the aquifer under a unit hydraulic gradient.

[2] The gradient is usually expressed as the difference in elevation of the water table between two points, divided by the distance between them. For example, a fall of one metre in water table elevation over a horizontal distance of 50 metres is a gradient of 1:50 (ie 0.02, expressed as a fraction).

Volume of groundwater flow

The groundwater flow through a unit area (eg one square metre) of an aquifer is determined by the aquifer hydraulic conductivity and the water table gradient, and is calculated from Darcy's Law: Flow = hydraulic conductivity x gradient.

Rate of groundwater travel

The rate at which groundwater travels through an aquifer is determined by the aquifer hydraulic conductivity, the water table gradient, and the aquifer porosity (expressed as a fraction). Rate of flow = hydraulic conductivity x gradient ∕ effective porosity[3].

Groundwater quality

Groundwater acquires soluble matter from the aquifer in which it is stored, and through which it moves. Generally, the longer the water remains in the aquifer, the more soluble constituents it acquires, and the poorer its quality. So, other things being equal, aquifers with relatively high hydraulic conductivity tend to have better quality water than low hydraulic conductivity aquifers. Also, other things being equal, better quality groundwater is found in aquifers in high rainfall areas, where groundwater recharges the aquifer more frequently, and aquifers are "flushed" more often.

In shallow unconfined aquifers, it is usual to find better quality groundwater near the water table where direct infiltration of rain has occurred. Quality typically decreases with depth.

A common measure of groundwater quality ('salinity') is its Total Dissolved Solids (TDS), expressed in milligrams per litre (mg/L; essentially the same as the older measure, parts per million, ppm). Typical TDS ranges of waters are:

	TDS (mg/L)
Tasmanian rain	<50
Tasmanian river water	<100
Drinking water starts to have 'taste'	250 – 500
Generally accepted desirable upper limit for drinking water	1,000
Range of commercially available mineral waters	100 – 1,500
Groundwater in coastal sands	450 – 800
Sea water	34,000

Hydrological cycle and groundwater systems

Figure 1 illustrates different components of the land-based part of the hydrological cycle[4] at the scale of a single catchment or smaller. The fundamentals of groundwater movement in an unconfined[5], fractured-rock, gravity-driven groundwater system are depicted schematically in the second figure.

In Figure 2, the hydraulic heads in the recharge areas are relatively high and decrease with depth. In discharge areas, the energy and flow conditions are reversed: heads are low and increase with depth. In between, the throughflow is almost horizontal as shown by the steeply dipping equipotential lines.

Figure 2 also illustrates the concept of a groundwater system[6]. In areas of moderate to high relief, the near-surface dominant groundwater flows to depths of the order of a hundred metres or so will

[3] For example, if the aquifer permeability is 2m^3/day/m^2, the gradient is 0.01 and the effective porosity is 0.1, the rate of flow would be 2 x 0.01 ∕ 0.1 = 0.2m/day.

[4] The *hydrological cycle* is the circulation of water in various phases through the atmosphere, over and under the earth's surface, to the oceans, and back to the atmosphere. The cycle is solar-powered. Because water is a solvent it dissolves elements, and geochemistry is a fundamental part of the cycle, which is a flux for water, energy, and chemicals.
Water enters the land-based cycle as precipitation; it leaves as surface streamflow (runoff) or evapotranspiration. The route which groundwater takes from a recharge point to a discharge point is a *flow path.*

[5] Locally (outcrop size or larger), the aquifer is probably confined by unjointed rock or clay weathering products or both. At increasing larger scales, the aquifer is unconfined.

[6] Sophocleous (2004) cited in Figure 2 defines a groundwater system as "a set of groundwater flow paths with common recharge and discharge areas. Flow systems are dependent on the hydrogeologic properties of the soil/rock material, and landscape position. Areas of steep or undulating relief tend to have dominant lo*cal flow systems* (discharging to nearby topographic lows such as ponds and streams). Areas of gently sloping or nearly flat relief tend to have dominant *regional flow*

be as local systems, with recharge on elevated areas discharging to streams. As depth increases, the dominant groundwater flows become increasingly intermediate and then regional in nature. It is therefore important to recognise the local site in the context of the larger groundwater system.

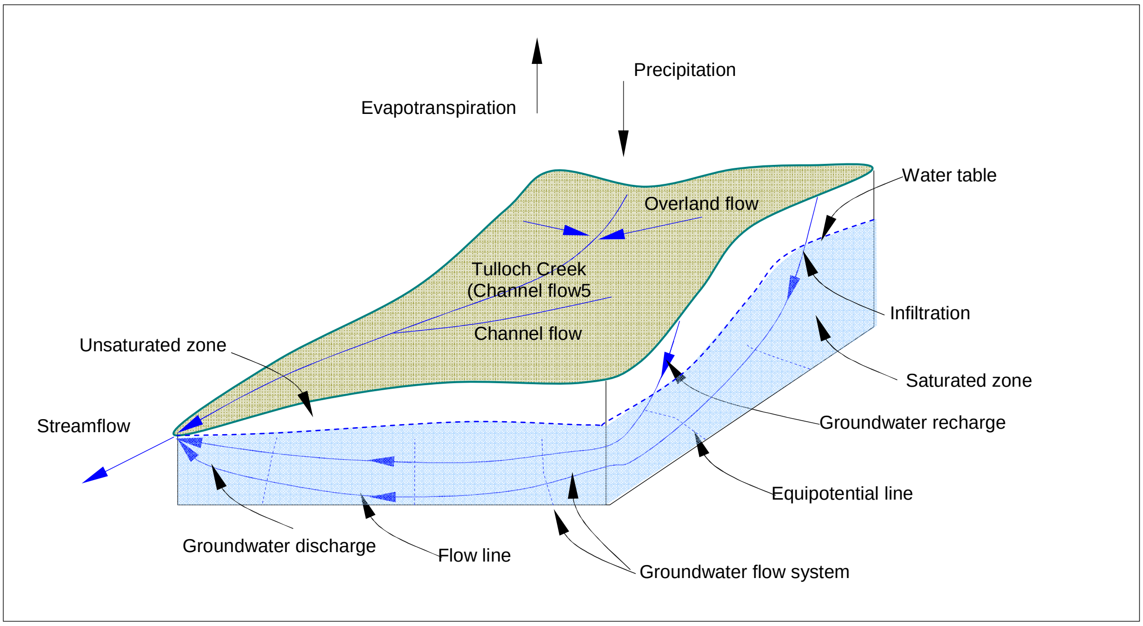


Figure 1 **Aspects of the land-based hydrological cycle.**

Hydrogeological characteristics of karst[7]

Karstic features like sinkholes, dolines, conduits, caves, sinking streams etc in soluble carbonate rocks may extend vertically to permanently, temporarily or ephemerally intersect the water table or piezometric surface in a fractured rock terrain. Joints, faults, bedding surfaces and other discontinuities in fractured rocks (carbonates or otherwise) constitute secondary porosity, karstic features constitute tertiary porosity.

Fluviokarst is a karst landscape "in which the dominant physical landforms are valleys initially cut by surface streams that have been partly or completely diverted underground by subsurface conduit piracy. This type of karst is often typified by carbonate rocks that have low intrinsic permeability..".

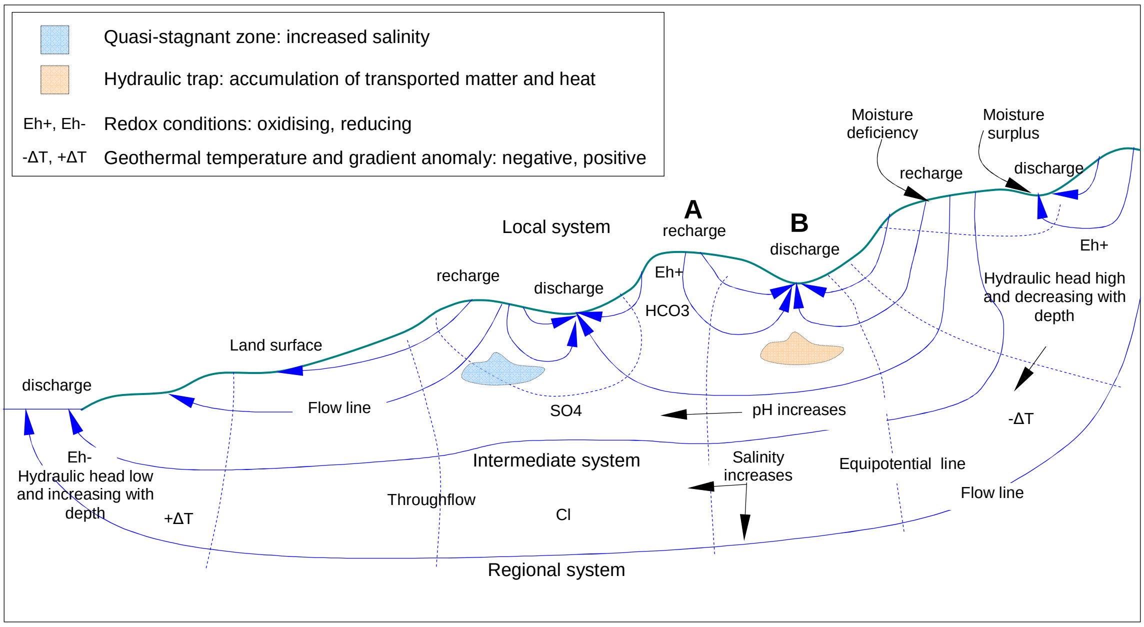
Doline karst is a karst landscape in which "surface streams are almost entirely absent, and almost all surface drainage is captured and drained internally by closed sinkhole depressions. This type of karst is typical of carbonate rocks that have high intrinsic permeability…", The physical and hydrologic distinctions between the two karst types are not always clearly defined, and many karst terranes have characteristics common to each.

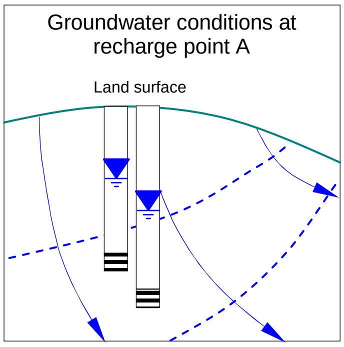
Conduits are open channels of any orientation formed in karst by the enlargement of existing secondary-porosity fractures, bedding surfaces and other defects. Conduits are tertiary-porosity features. A distinctive feature of karst aquifers are the "typically dendritic or branching networks of conduits that meander among bedding units, join together as tributaries, and increase in size and order in the downstream direction" (Figure 7.4). In the simplest terms, these conduit networks grow by way of a complex hydraulic-and-chemical feedback loop, in which the basic steps are: conduit growth and enlargement → increased hydraulic capacity → increased discharge → enhanced dissolution and physical corrosion → additional conduit enlargement → subsurface piracy of flows in smaller conduits by the larger conduits. In this process, the largest conduits act as master drains that locally alter the hydraulic flow (or equipotential) field so as to capture ground water from the

systems (discharging at much greater distances than local systems in major topographic lows or oceans)." A three-dimensional closed groundwater flow system that contains all the flow paths is called the *groundwater basin*.

[7]This Section draws on Taylor, C. J. and Greene, E. A. (2008). Hydrogeologic Characterisation and Methods Used in the Investigation of Karst Hydrology. [Chapter 3 of Rosenberry, D. O. and LaBaugh, J. W. (2008). Field Techniques for Estimating Water Fluxes Between Surface Water and Groundwater. Techniques and Methods 4 – D2. US Geological Survey]. Some text is quoted.

surrounding aquifer matrix, the adjoining fractures, and the smaller nearby conduits. Depending on their sizes (hydraulic capacity) and organization (interconnection), conduit networks are capable of discharging large volumes of water and sediment rapidly through a karst aquifer. Flow velocities in well-developed and well-integrated conduit networks up to hundreds to thousands of metres per day are not uncommon.





In recharge areas (at left), the hydraulic heads are relatively high and decrease with depth, as shown by the water levels in two adjacent piezometers. In discharge areas (at right), the energy and flow conditions are reversed: heads are low and increase with depth.

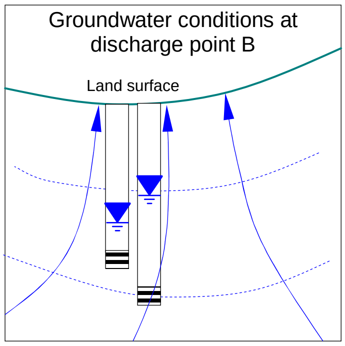


Figure 2 **Fundamentals of groundwater hydrology in a gravity-driven groundwater system.** Adapted from Sophocleous (2004). Groundwater recharge, in *Groundwater*, [Eds. Luis Silveira, Stefan Wohnlich and Eduardo J. Usunoff] in *Encyclopaedia of Life Support Systems* (EOLSS), *Developed under the Auspices of the UNESCO, Eolss Publishers, Oxford, UK, [www.eolss.net)*

Karst **springs** are the natural outlets for water discharging from conduit networks. They typically are developed at a local or regional ground-water discharge boundary—that is, at a location of minimum hydraulic head in the aquifer—often at or near the elevation of a nearby base-level surface stream."

Recharge in karst

Karst loss is surface runoff diverted to subsurface karst features via autogenic or allogenic diffuse or concentrated recharge. Sometimes, the karst features may not be obvious at the surface but karst loss still occurs. Evidence might include misfit surface streams, and disappearing streams.

Estimating karst loss in hydrological modelling of karst terrain usually requires field inspection and mapping of karstic features, and a structured approach (Anon, 1995?):

> "1. Delineate the contributing drainage area or watershed to be studied.
> 2. Define any sinkhole areas within the contributing drainage area where surface drainage has no means of escaping offsite, other than downward through the karst strata (i.e. cracks, sinks, etc.). These areas can be assumed to contribute no surface discharge and can be subtracted from the contributing drainage area from Step 1.

3. Determine the amount of the contributing drainage area (from Step 2) underlain by karst strata (in percent).
4. Calculate the peak rate of runoff from the contributing drainage area using standard hydrologic methods[8], and reduce the calculated value by multiplying by the *Karst Loss Modification Value* based on the percent karst (% Karst) calculated in Step 3."

See Table 1 for Karst Loss Modification Values.

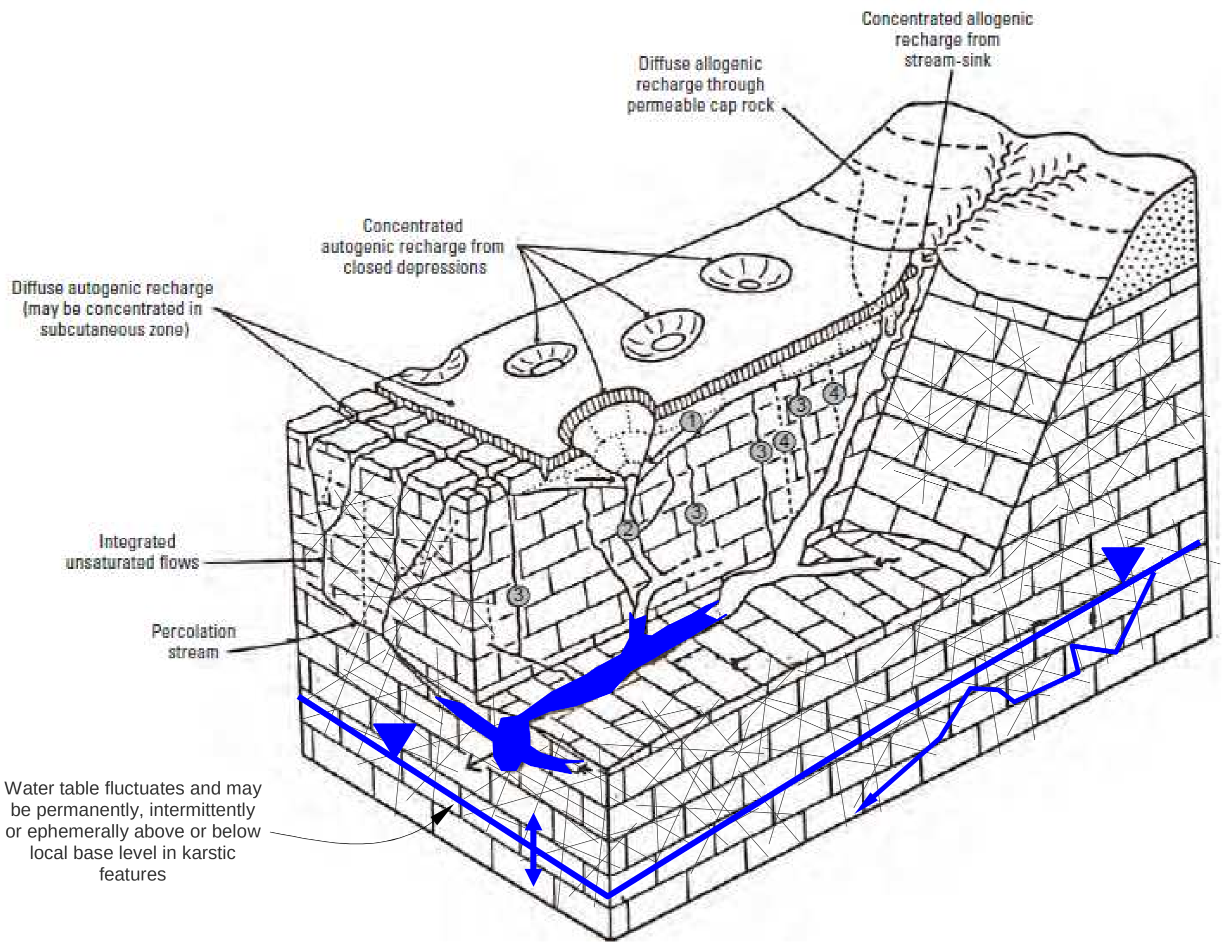


Figure 3. **Hydrogeological block diagram showing diffuse and concentrated allogenic and autogenic karstic recharge.**
Autogenic recharge originates as precipitation falling directly on karstic rocks. Allogenic recharge originates as precipitation falling on non-karstic rocks. Water flows through the unsaturated zone via (1) diffuse flow through soil or unconsolidated surface materials, (2) concentrated flow through solution-enlarged sinkhole drains, (3) diffuse infiltration through vertical fractures, and (4) diffuse infiltration through permeable rock matrix. Open conduits shown as solid blue are filled with ground water.
Modified from Taylor, C. J. and Greene, E. A. (2008) cited earlier.

[8] For example, the Rational Method.

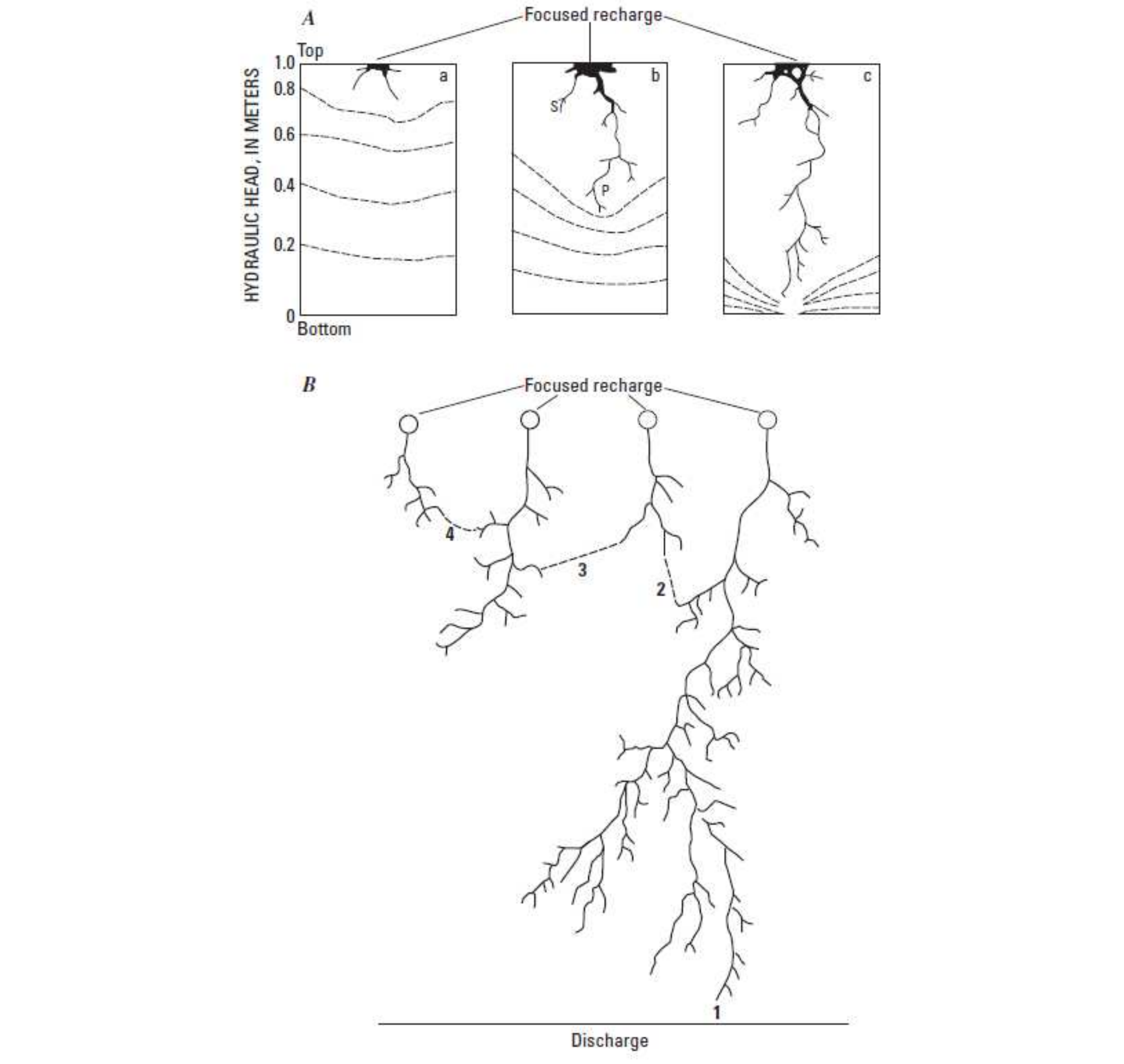


Figure 4 **Competitive growth of conduits and distortion of hydraulic flow fields**

A (a) Initiation of recharge
(b) Change in hydraulic gradient in response to faster growing primary (P) conduit and secondary (S) conduit
(c) Primary conduit breaks through to discharge boundary, slowing or inhibiting growth of secondary conduit.

B Sequence of development of integrated drainage network due to faster growth and breakthrough of primary conduit (1) and subsequent capture of flow and linking of secondary conduits (2 – 4)

[Figure 2 of Taylor and Greene (2008) cited earlier]

Table 1 **Karst Loss Modification Values**

Source: Anon (1995?): http://www.deq.virginia.gov/Portals/0/DEQ/Water/Publications/TechBulletin2.pdf

% Karst	Storm Return Frequency (years) 2	10	100
100	0.33	0.43	0.50
90	0.35	0.46	0.56
80	0.38	0.51	0.62
70	0.47	0.58	0.68
60	0.55	0.66	0.74
50	0.64	0.73	0.80
40	0.73	0.80	0.85
30	0.82	0.86	0.89
20	0.91	0.92	0.93
10	1.00	0.98	0.97
0	1.00	1.00	1.00