# Groundwater Systems—A Hydrogeological Typology

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## Glossary

**Aquifer** Water-bearing units underground capable of yielding water to wells or springs.
**Confined aquifer** Aquifer where the groundwater is pressurized such that it rises above the bottom of an overlying confining unit when the aquifer is penetrated by a well.
**Confining unit** Geological units significantly less permeable than aquifers.
**Consolidated rocks** Rocks in which the individual minerals are firmly held together; includes sedimentary rocks where grains are cemented as well as igneous rocks (formed by crystallization from magma, e.g., basalt or granite) and metamorphic rocks (formed under high temperature and pressure, e.g., gneiss).
**Groundwater** Water occupying the pore spaces below the (subsurface) water table.
**Piezometric surface** Imaginary surface to which groundwater rises if a confined aquifer is penetrated by a well.
**Primary porosity** Pores created when the rock was originally formed.
**Secondary porosity** Porosity developed after the original formation of the rock; more narrowly defined, discontinuities (fractures) separating blocks of intact rock, which were created by mechanisms other than dissolution.
**Tertiary porosity** Conduit porosity developed where the dissolution of rock widened fractures.
**Unconfined aquifer** Aquifer where the water table is located within the water-bearing unit (water-table aquifer).
**Unconsolidated rocks** Sediments composed of loosely arranged particles or grains (not cemented).
**Water table** Surface where the water pressure is equal to the atmospheric pressure.

## Introduction

Groundwater is a major water resource for the world population and sustains streams, lakes, and wetlands. Sustainable use and protection of groundwater must be based on a sound understanding of groundwater flow and storage. As part of the hydrological cycle, groundwater systems are dependent on hydrological processes controlling their recharge and discharge. Yet, regional and local assessments of groundwater systems also require knowledge of subsurface properties controlling groundwater flow and storage. The aim of this chapter is to introduce the basic terms and concepts needed to characterize and classify groundwater systems with regard to these hydrogeological properties and their effect on flow and storage.

The following section introduces basic definitions of different types of subsurface waters, in particular groundwater. Subsequently, different types of aquifers are distinguished based on the hydraulic conditions found in the subsurface. Then different types of porosity are considered, which correspond to a classification of aquifers based on the geological material. Building on this typology, aquifer properties controlling groundwater flow and storage are introduced and discussed with regard to the different aquifer types. Finally, chemical and thermal properties of groundwater and how they are interlinked with groundwater flow are discussed.

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## Basic definitions

After intense rainfall or snowmelt, the soil or rock at the ground surface may temporarily be saturated with water. Generally, however, the shallow subsurface contains both air and water and thus is termed the *unsaturated zone* (or vadose zone) (Fig. 1). Underneath the unsaturated zone follows the *saturated zone* (or phreatic zone), where all voids are filled with water. If observation wells are drilled into the saturated zone, the lower part of the well fills with water. The top of the water column in the well marks the *water table*. Water below the water table is termed *groundwater*. At the water table, the water pressure is equal to atmospheric pressure. Below the water table, the water pressure increases with increasing depth. Conversely, the pressure in the unsaturated zone is lower than atmospheric pressure. This is a result of the capillary forces within narrow pores spaces and leads to a *capillary fringe* where water is held above the water table.

Processes within the unsaturated zone and the capillary fringe, such as the infiltration and downward percolation of water, play an important role in replenishment of groundwater systems and in the fate of contaminants released at the ground surface or shallow subsurface. Yet, these processes are dealt with elsewhere. The remainder of this chapter focusses on the zone below the water table, i.e., groundwater.

## Types of aquifers

*Aquifers* are water-bearing units capable of yielding water to wells or springs; the unsaturated zone of such a unit is understood as being part of the aquifer (Todd and Mays, 2005). Aquifers can be overlain or underlain by *confining units* that are significantly less permeable than the aquifer. Confining units can be further classified into aquicludes (containing but not transmitting water), aquifuges (neither containing nor transmitting water), and aquitards (transmitting less water than adjacent aquifers). Evidently, aquifer and confining unit are relative terms. A geological unit that yields only little water may be considered an aquifer if other units within the same region are even less productive, whereas the same geological unit may be classified as confining unit if overlain or underlain by rocks that yield more water. A quantitative measure of the capability to transmit water thus is needed and will be introduced in the section "Flow properties".

Based on the position of the water table different types of aquifers can be distinguished (Fig. 2). In an *unconfined aquifer* (or water-table aquifer) the water table is located within the water-bearing unit (upper aquifer in Fig. 2). This condition is often found where permeable rock crops out at the land surface. A *perched aquifer* is a special case of an unconfined aquifer that occurs if a local groundwater body develops on top of a discontinuous confining layer above the regionally extensive water table of a main groundwater body. Thus, the perched aquifer and the underlying groundwater body are separated by an unsaturated zone. The groundwater within a *confined aquifer* is pressurized such that it rises above the bottom of an overlying confining unit when the aquifer is penetrated by a well (lower aquifer in Fig. 2). The imaginary surface constructed from water levels of such wells is termed

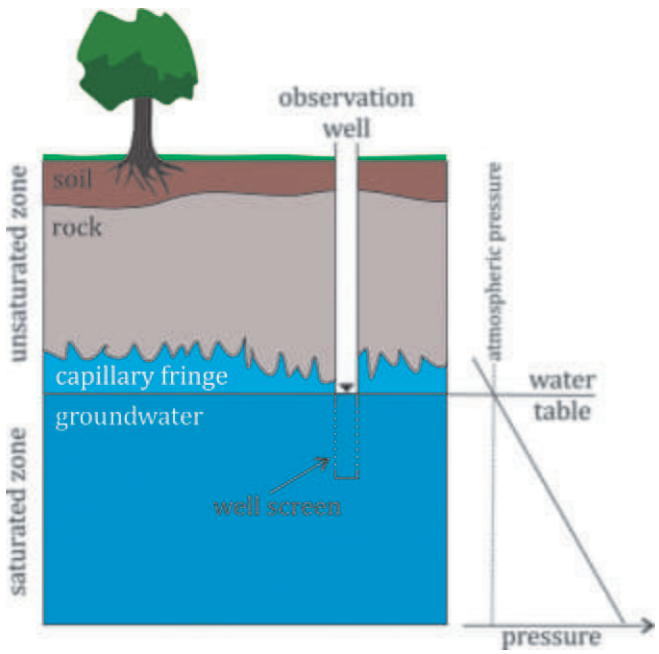


**Fig. 1** Zonation of the subsurface and basic hydrogeological terms. Groundwater enters the well through the slotted part of the well casing termed well screen (dashed line). The water table observed in the well is marked by an inverted triangle.

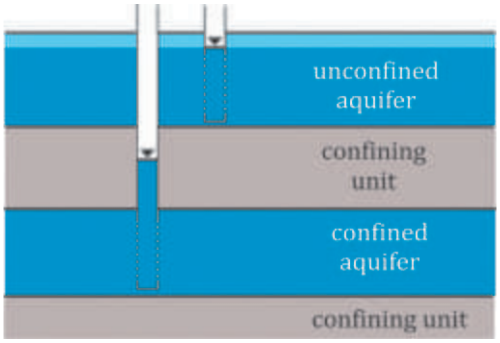


**Fig. 2** Definition of unconfined and confined aquifers based on the relative position of the water table observed in the well.

*piezometric surface* (or potentiometric surface). If the pressure is sufficiently high to rise the water above ground level a flowing artesian well results. Confined or artesian aquifers may be found when a water-bearing unit crops out in a topographically elevated recharge area and dips into a basin where it is covered by a confining unit.

Distinguishing between unconfined and confined aquifers is important in several ways. From a practical point of view, it is vital to recognize confined conditions when drilling into the underground, as the rise of groundwater in the borehole might require technical measures that need to be planned in advance. Evidently, this particularly applies to artesian conditions where the water rises above ground level. Yet, also groundwater flow from a confined aquifer into an overlying unit generally results in undesired changes of hydraulic conditions and potentially adverse effects e.g., on water quality. A striking example is a case reported from Staufen, Germany, where confined groundwater had been allowed to rise in a borehole and caused swelling of the adjacent rock; the resulting uplift of the land surface caused severe damages to buildings (Sass and Burbaum, 2010). Confined and unconfined conditions also need to be distinguished from the more theoretical perspective of groundwater hydraulics. Although comprehensive treatment of groundwater hydraulics is beyond the scope of this chapter, fundamental differences in the storage properties of confined and unconfined aquifers are addressed below in the section "Storage properties".

## Types of porosity

The occurrence of groundwater requires voids or pore spaces in the underground that can take up water. The percentage of pore spaces in a rock volume is termed (total) *porosity*. Different geological materials not only have different values of porosity; more importantly, also the size, shape, and connectivity of the pore spaces differ. The different nature of the pore spaces is closely related to their origin and thus to the type of geological material. Hence, hydrogeologists distinguish different types of porosity depending on when and how the pore spaces where formed.

*Primary porosity* denotes pores created when the rock was originally formed. This type of porosity is most relevant in unconsolidated sediments. Such sediments consist of grains that were transported and deposited by wind, water, or ice. The size distribution of these grains is one major factor controlling porosity. As a rule, total porosity is higher the smaller and the more uniform the grain sizes are (Fig. 3 and Table 1). Sediments transported by wind tend to be well sorted, i.e., display a narrow range of grain sizes. Since the transport and deposition of sediments by water is controlled by characteristic relationships between grain size and flow velocity, one may also expect sediments deposited in river channels and floodplains to be well sorted. However, flow velocities in such environments may vary considerably in space and time. Across the river valley, the grain size distribution of such sediments therefore may exhibit significant variability. Sediments deposited from ice generally are expected to be poorly sorted and thus may display a wide range of grain sizes. The shape and packing of the grains are further factors affecting the porosity of unconsolidated sediments. Compaction by overlying rock and cementation by chemical precipitates transforms the unconsolidated sediments to consolidated sedimentary rock. Primary porosity is reduced by this process of lithification, which is why secondary porosity gains importance in this type of rock.

*Secondary porosity* generally denotes porosity developed after the original formation of the rock and comprises all types of discontinuities separating intact blocks of consolidated rocks. We distinguish fracture porosity, which includes discontinuities formed by a variety of processes, such as lithostatic or fluid pressures, tectonic forces, and thermal effects, from conduit porosity, which develops where the dissolution of rock widens fractures (Fig. 4). Consequently, conduit porosity is sometimes referred to as *tertiary porosity* (Ford and Williams, 2007). Rocks that are sufficiently soluble to enable the development of conduit porosity are termed karst rocks and, in particular, include carbonate rocks (e.g., limestone) and evaporite rocks (e.g., gypsum). It is important to note that several types of porosity can occur simultaneously. For example, sedimentary rocks such as sandstone can be fractured but still may have significant primary porosity. Likewise, dissolution may create conduits in carbonate rocks, while the majority of fractures may be little affected by dissolution and primary porosity may exist as well.

It is common to name aquifers after the geological material or the type of porosity primarily enabling groundwater flow. Hence, permeable unconsolidated sediments, such as sand and gravel, where groundwater is transmitted through primary porosity, are

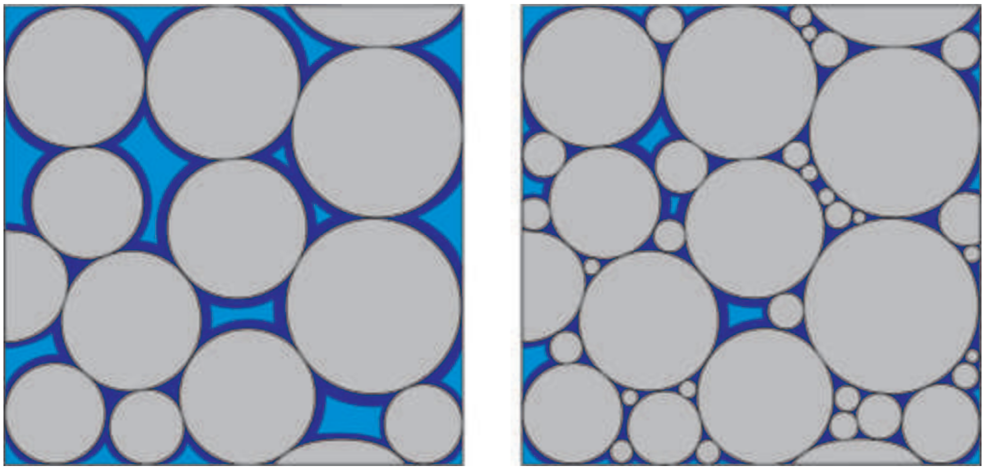

**Fig. 3** Schematic sketch of the primary porosity (blue colors) of an unconsolidated sediment. The relatively coarse and well sorted sediment on the left has a higher total porosity than the poorly sorted sediment on the right, which contains fine grains filling spaces between the coarse grains. Dark blue color indicates immobile water surrounding the grains; only the light blue colored part of the pore space is available for groundwater flow (effective porosity).

**Table 1** Approximate ranges of total porosity, effective porosity, and hydraulic conductivity for selected geological materials.

| *Geological material* | | *Total porosity*[a] *(%)* | *Effective porosity*[a] *(%)* | *Hydraulic conductivity*[b] *(m $s^{-1}$)* |
|---|---|---|---|---|
| Unconsolidated rocks | Gravel | 28–34 | 15–30 | $10^{-3}$–1 |
| | Sand | 35–50 | 10–30 | $10^{-6}$–$10^{-2}$ |
| | Silty sand | 33–40 | 8–12 | $10^{-7}$–$10^{-3}$ |
| | Silt | 40–50 | 5–20 | $10^{-9}$–$10^{-5}$ |
| | Clay | 40–60 | 1–5 | $10^{-13}$–$10^{-9}$ |
| Consolidated rocks | Karstified limestone | 10–25 | 0.5–10 | $10^{-6}$–$10^{-2}$ |
| | Fractured basalt | 5–30 | 2–10 | $10^{-7}$–$10^{-2}$ |
| | Sandstone | 15–30 | 5–25 | $10^{-10}$–$10^{-6}$ |
| | Unfractured igneous or metamorphic rock | 0–5 | 0–3 | $10^{-13}$–$10^{-10}$ |

[a]From Singhal BBS and Gupta RP (2010) Applied Hydrogeology of Fractured Rocks (2nd edn.). Dordrecht: Springer; Hölting B and Coldewey WG (2019) *Hydrogeology*. Berlin: Springer for silty sand.
[b]From Freeze RA and Cherry JA (1979) *Groundwater*. Englewood Cliffs, NJ: Prentice-Hall.

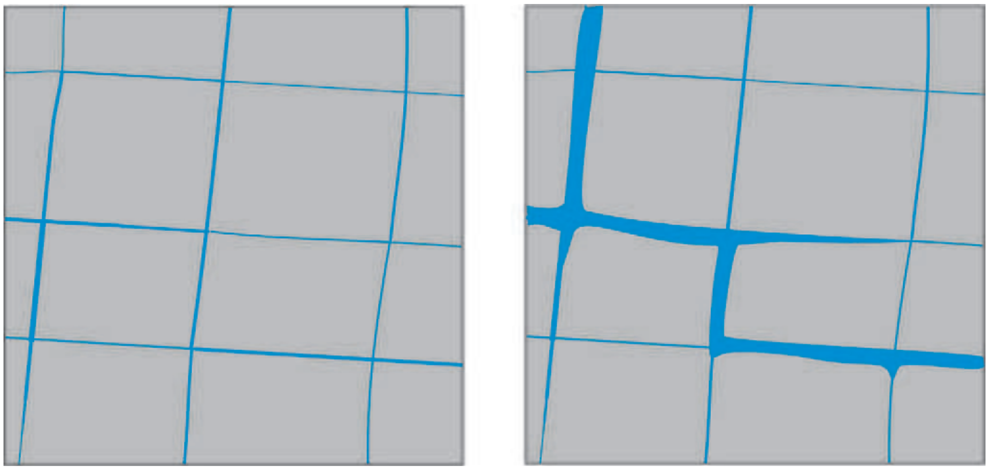

**Fig. 4** Schematic sketch of fracture porosity resulting from discontinuities separating intact blocks of consolidated rocks (left). Conduit porosity develops where fractures are widened by dissolution of rock (right).

termed *unconsolidated aquifers* or *porous aquifers*. Correspondingly, the term *consolidated aquifer* refers to all types of permeable consolidated rocks. Commonly, these rocks are fractured and thus the aquifers are termed *fractured aquifers*. Some consolidated rocks, for example sandstone and basalt, the latter of which originates from volcanic lava flows, additionally may have primary porosity. Nevertheless, they usually will be classified as fractured aquifers, as flow typically occurs through open fractures rather than through the narrower pore spaces provided by the primary porosity. Similarly, solution conduits, which commonly have openings ranging from centimeters to meters, are expected to be more effective in conducting water than the pores or fractures of karst rocks. Aquifers where conduit porosity controls groundwater flow thus are termed *karst aquifers* or *karstified aquifers*.

The transmission of groundwater requires interconnected pore spaces of sufficient size to overcome adhesive forces attracting the water molecules to the mineral surfaces. The volume percentage of such pore spaces is termed *effective porosity*. The pore size in unconsolidated sediments tends to decrease with decreasing grain size. As a consequence, large parts of the pore space of fine-grained materials, such as clay or silt, is occupied with immobile water (Fig. 3). Hence, these materials have effective porosities considerably lower than their total porosity (Table 1). As a consequence, clay or silt are hardly capable of transmitting water and are classified as confining units rather than aquifers. Coarse materials such as sand and gravel, particularly if well sorted, tend to have high effective porosity that might come close to their total porosity (Table 1). Such materials represent aquifers, even though their total porosity generally is lower than that of fine-grained materials. The capability of rocks to transmit water thus appears to be related to their effective porosity rather than their total porosity. In this regard, the major difference between the two sediments sketched in Fig. 3 is not so much their different total porosity, but that the coarse and well sorted material has larger openings and thus higher effective porosity than the poorly sorted sediment. Yet, particularly some consolidated rocks may have low total porosity and consequently low effective porosity, but if few interconnected fractures or conduits have large openings these rocks are still able to transmit water. A more quantitative measure of aquifer permeability thus is needed and introduced below.

## Flow properties

With some exceptions, such as flow through highly conductive conduits or flow in the vicinity of highly productive pumping wells, groundwater flow is very slow, in the order of meters per day or even well below that. Under such conditions, the *specific discharge q*, i.e., the volumetric rate of flow per unit area of flow cross section, is found to be proportional to the *hydraulic gradient i*, the slope of the water table (or piezometric surface). This is defined by Darcy's law (Darcy, 1856)

$$q = K\,i \tag{1}$$

where the proportionality constant $K$ is termed *hydraulic conductivity*. Hydraulic conductivity estimates can be obtained by solving Darcy's law for $K$ and inserting values of specific discharge and hydraulic gradient measured in lab or field experiments.

Hydrogeologist often treat hydraulic conductivity as if it were a property of the rock. Strictly, this is not correct, because $K$ is also dependent on the density and viscosity of the fluid. Yet, in many groundwater systems the aforementioned fluid properties are spatially and temporally nearly constant, such that variations in $K$ almost entirely result from differences in geological materials. Assuming constant fluid properties, values of $K$ for different materials vary by many orders of magnitude (Table 1). Hydraulic conductivities of unconsolidated aquifers such as sand and gravel typically range from $10^{-5}$ to $10^{-2}$ m s$^{-1}$, though lower or higher values are possible. Values of confining units, e.g., composed of silty or clayey sediments, are orders of magnitude lower. Similarly, consolidated rocks exhibit a wide range of hydraulic conductivity depending on the characteristics of their fracture or conduit porosity.

Assessing the hydraulic conductivity of rocks is complicated by their *heterogeneity* and *anisotropy*. The grain size distribution of unconsolidated sediments and the frequency as well as the size of fractures or solution conduits of consolidated rocks commonly vary in space. This heterogeneity is displayed in the spatial distribution of hydraulic conductivity values. As a consequence, measurements of $K$ at different locations may yield very different results even within the same geological setting. As a further consequence, hydraulic conductivities obtained by lab or field experiments depend on the scale of observation. This is most obvious for consolidated rocks with several types of porosity, such as sandstone or limestone. Lab measurements conducted at small samples of such rocks mainly provide information about the hydraulic conductivity associated with the primary porosity, whereas the effects of fracture or conduit porosity can only be observed at large spatial scales. In addition, sediment structures, fractures, and conduits may have preferential directions. This leads to a directional dependence, i.e., anisotropy, of hydraulic conductivity. As an example, the hydraulic conductivity of unconsolidated sediments composed of multiple layers generally is higher in the horizontal than in the vertical direction. Consolidated rocks tend to exhibit even higher anisotropy of hydraulic conductivity, because the mechanisms creating fractures and conduits often favor distinct directions.

Characterizing an aquifer's ability to transmit water is further complicated when groundwater flow is fast. Darcy's law holds only if inertial forces and turbulent friction can be neglected, which is not met if the flow velocity is too high. This is generally expected in karst aquifers, where velocities are often in the order of kilometers per day, and also occurs in other highly conductive aquifers close to productive pumping wells. The additional forces increase the flow resistance and cause deviation from the linear relationship between hydraulic gradient $i$ and specific discharge $q$ stated by Darcy's law. One approach to account for this effect is the non-linear flow equation proposed by Forchheimer (1901).

$$i = a\,q + b\,q^2 \tag{2}$$

where $a = 1/K$ is the reciprocal of hydraulic conductivity and the Forchheimer coefficient $b$ an aquifer property controlling the non-linear deviation from Darcy's law. Estimates of $b$ for various geological materials are provided by Zeng and Grigg (2006), Chin et al. (2009), and Sidiropoulou et al. (2009). Compared with hydraulic conductivity values of rocks, the non-linear flow properties of geological materials have been considerably less investigated though.

The ability of an aquifer to transmit water, characterized by its hydraulic conductivity $K$ (and under the above specified conditions the Forchheimer coefficient $b$), is one important control on the interaction of groundwater systems with other components of the water cycle. A highly conductive aquifer is able to transmit the entire recharge under a hydraulic gradient much lower than the slope of the land surface (Fig. 5, left). In such recharge-controlled systems, the water table of the elevated areas is at great depth and thus disconnected from the hydrological processes at the land surface. In contrast, the transmission of groundwater through lowly conductive rocks involves steep hydraulic gradients such that the water tables of the elevated areas are shallow and thus interacting with the evapotranspiration and runoff processes at the land surface (Fig. 5, right). In such topographically-controlled systems, the potential recharge exceeds the rate at which groundwater can be laterally transmitted by the aquifer, i.e., recharge is rejected (Theis, 1940; Cuthbert et al., 2019).

## Storage properties

In a hypothetical groundwater system where recharge and discharge are in balance, the stored water volume remains constant in time. However, natural or artificial fluctuations in recharge and discharge commonly cause temporal changes in the stored water volume. For example, consider the situation where pumping is started in an unconfined aquifer. As pumping starts, water is withdrawn from the wellbore, causing the water level in the pumping well to drop. Continued pumping then requires that groundwater flows from the aquifer towards the well. As a result, groundwater is drained from the aquifer storage and the water table in the vicinity of the pumping well starts to decline (Fig. 6, left). Hence, water withdrawal from the aquifer storage involves changes in water levels. To characterize this relationship quantitatively, we define the *storativity* (or storage coefficient) of an aquifer as the water volume released from storage per unit surface area per unit decline in water level.

In the above example of an unconfined aquifer, the water is released by gravity drainage from the aquifer storage. The water volume per rock volume released by gravity drainage is termed *specific yield*; it is equivalent to the storativity of the unconfined aquifer. Since some percentage of the water is retained as thin films surrounding grains or as capillary water in narrow pores, the specific yield is lower than the total porosity. The approximate ranges for effective porosity given in Table 1 also apply to specific yield (Singhal and Gupta, 2010). Thus, the specific yield of unconsolidated sand or gravel roughly ranges between 0.1 and 0.3, while values of fined-grained sediments are lower. Highly fractured, consolidated rocks may reach values similar to those of

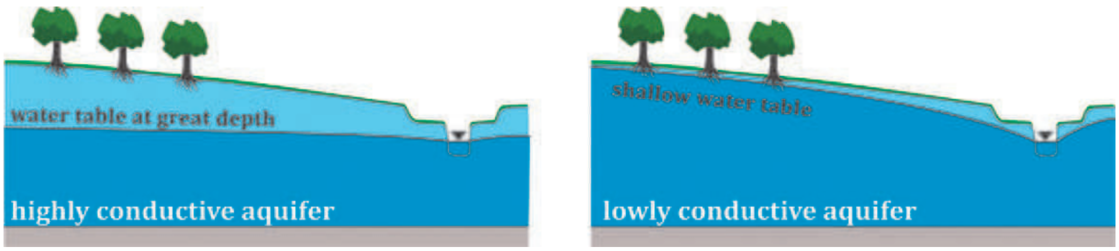


**Fig. 5** The water table of a highly conductive aquifer (left) tends be at great depth below the land surface, whereas the transmission of groundwater through a lowly conductive rock requires a steep hydraulic gradient, thus leading to shallower water tables (right).

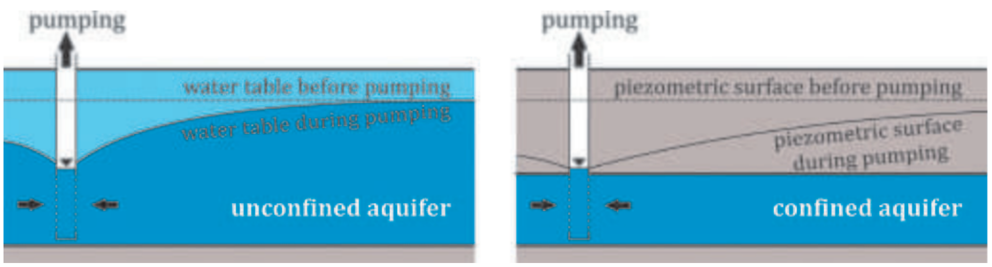


**Fig. 6** Comparison of unconfined and confined aquifers when they are pumped. In the case of the unconfined aquifer (left) water is released by gravity drainage from the aquifer volume enclosed by the water tables before and during pumping. In contrast, the pore spaces of the confined aquifer (right) remain saturated during pumping; water is only released from the aquifer storage because the lowering of the piezometric surface corresponds to a reduction of pressure and thus leads to expansion of water and reduction of pore space.

coarse-grained unconsolidated rocks, but often the specific yield of consolidated aquifers is lower than that of unconsolidated aquifers.

Consider now a confined aquifer that is pumped (Fig. 6, right). As long as the aquifer is confined, the pore space remains saturated despite the lowering of the water level in the pumped well. Nevertheless, groundwater is released from the aquifer storage, because the decline of pressure, which corresponds to the lowering of the piezometric surface, results in an expansion of water and compression of the rock and thus the reduction of pore space due to the load of the overburden. The water volume that can be released by this mechanism and thus the storativity of confined aquifers can be estimated from the compressibilities of rock and water. The storativity of confined aquifers typically ranges in the order of $10^{-3}$ to $10^{-5}$. This is much lower than typical values of specific yield and indicates that in unconfined aquifers the described effect of decompression generally is negligible compared with gravity drainage.

The greatly different values of storativity constitute an important difference between confined and unconfined aquifers. This can be understood by considering a pumping situation as described above (Fig. 6). Since the same decline in water level releases much less water per unit volume from a confined than from an unconfined aquifer, the confined aquifer requires a much larger aquifer volume to provide the same amount of water. As a consequence, after a given time the area affected by pumping will be larger for the confined than for the unconfined aquifer (assuming they have the same hydraulic conductivity). Similarly, disturbances by flood pulses or effects of droughts propagate faster (or over a greater distance in a given time) in a confined aquifer than in an unconfined aquifer of otherwise similar properties.

## Chemical and thermal properties

Groundwater systems can also be characterized and classified based on their chemical and thermal properties. A comprehensive treatment of this topic is beyond the scope of this chapter. Yet, basic concepts and their interrelation with groundwater flow are briefly introduced below.

The water infiltrating into the soil after rainfall or snowmelt initially has a low solute concentration. As the water moves through the soil, it dissolves gases such as carbon dioxide and reacts with the soil organic matter and minerals. Thus, the water is already enriched in solutes when it reaches the water table. Depending on the geochemical conditions in the aquifer, the chemical composition of the groundwater further evolves through time.

The solute content of groundwaters reflects the processes involved in the hydrogeochemical evolution along the flow path. Thus, groundwaters can be classified into various types of hydrochemical facies based on the major ions they contain (Back, 1966). The cation facies can be characterized as calcium type, magnesium type, sodium and potassium type, or no dominant type; the anion facies can be classified as bicarbonate type, sulfate type, chloride type, or no dominant type (Fig. 7). The chemical composition of the geological material evidently is a major control on the hydrochemical facies. For example, limestone is composed of calcium carbonate and thus groundwater flowing through limestone that creates conduit porosity by dissolving the rock is expected to be of calcium-bicarbonate type. If water originating from limestone penetrates gypsum, which is composed of calcium sulfate, the increasing load of calcium leads to chemical precipitation of calcium carbonate and the hydrochemical facies changes towards calcium-sulfate type. Similar processes occur in other rocks such that the chemical composition evolves with increasing travel time and distance along the groundwater flow path. Generally, the content of dissolved solids increases with increasing distance and depth of the groundwater circulation, and the anionic facies is expected to change from bicarbonate to sulfate and chloride type, i.e., towards anions originating from highly soluble minerals (Tóth, 1999).

The chemical evolution of groundwaters is also influenced by thermal conditions within the aquifers, as changes in water temperature may disturb chemical equilibria and thus enable further dissolution or chemical precipitation of minerals. The surficial zone of the subsurface is influenced by seasonal temperature fluctuations, the amplitude of which decreases with depth (Anderson, 2005). At a depth of around 10–20 m below ground level, seasonal fluctuations are usually no longer observed (Todd and Mays, 2005), and thus the shallow groundwater exhibits temperatures close to the mean annual air temperature. In the geothermal zone that follows below, the temperature of the rock increases with depth according to the geothermal gradient (on average three Kelvin per 100 m). Since slowly circulating groundwater is in thermal equilibrium with the rock, the groundwater temperature increases with depth too. As a consequence, deep groundwater systems may exhibit high groundwater temperatures and thus constitute thermal water resources.

Changes in solution content and temperature involved in the deep circulation of groundwaters affect the density of the groundwater and thus potentially are able to induce density-driven flow processes. Moreover, the viscosity of water decreases with increasing temperature, which affects the hydraulic conductivity (see section "Flow properties"). Thus, potential changes in the density and viscosity of the water resulting from the chemical and thermal evolution along the flow path need to be considered in the assessment of deep groundwater systems. Similar effects can be found in coastal aquifers where freshwater and saltwater interact or in groundwater systems where thermal impacts resulting from human activities affect the density or viscosity of the water.

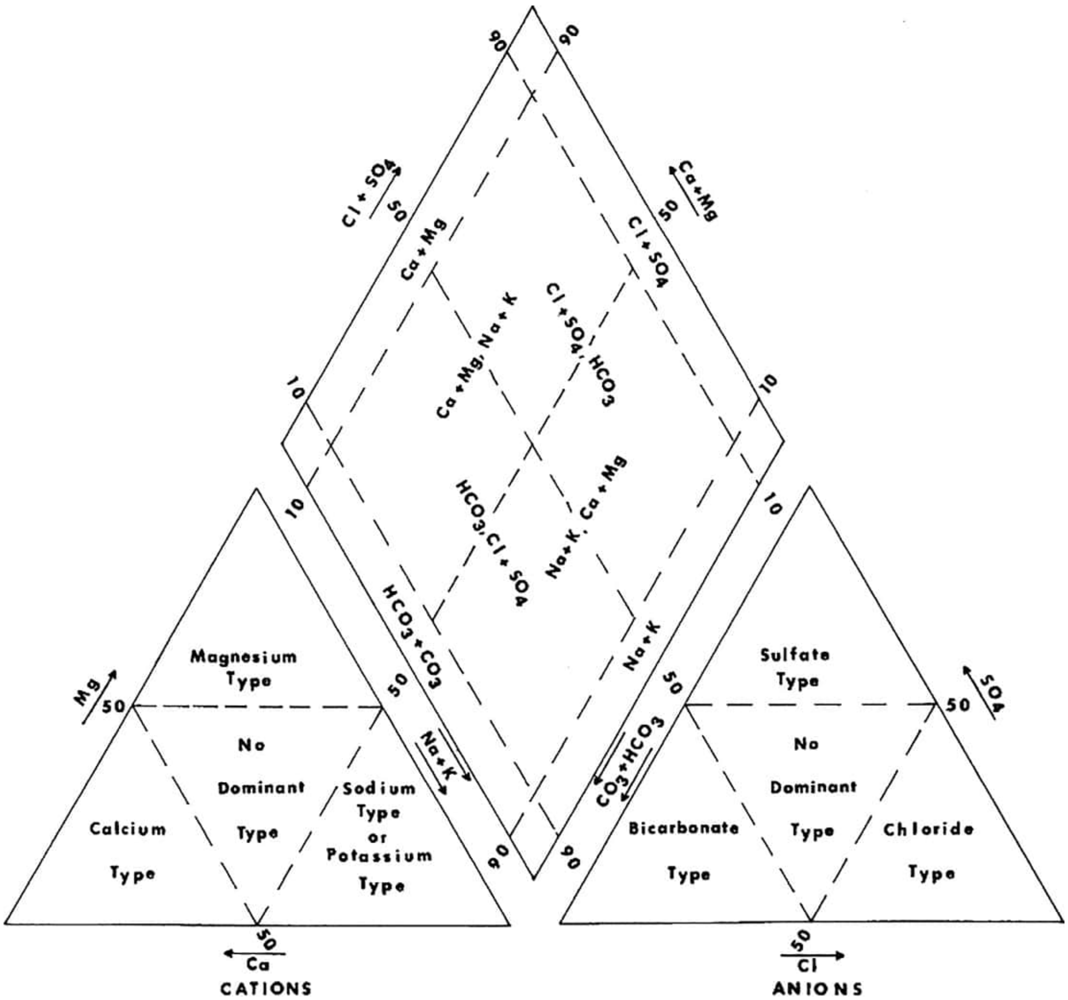


**Fig. 7** Piper diagram showing hydrochemical facies. Numbers indicate percentages of major ion species based on ion concentrations in milliequivalents per liter (Ca = calcium, Mg = magnesium, Na = sodium, K = potassium, $CO_3$ + $HCO_3$ = bicarbonate, $SO_4$ = sulfate, Cl = chloride). From Back W (1966). *Hydrochemical Facies and Ground-Water Flow Patterns in Northern Part of Atlantic Coastal Plain. Geological Survey Professional Paper 498-A.* Courtesy the U.S. Geological Survey.

## Conclusion

The hydrogeological typology of groundwater systems rests on two pillars, the distinction between unconfined and confined aquifers as well as the differentiation of porosity types. The first is defined by the position of the water table relative to the hydrogeological units; the second is determined by the characteristics of the geological material. The framework provided by this typology supports the understanding of aquifer properties governing flow and storage processes in groundwater systems. Groundwater systems can be further classified based on other aspects such as the chemical or thermal properties of the water. Possible interrelations between groundwater flow, chemistry and temperature need to be considered particularly in the hydrogeological assessment of deep groundwater systems, coastal aquifers, or settings affected by human impacts.

## Knowledge gaps

- How to deal with scale effects resulting from the occurrence of several types of porosity in hydrogeological assessments of consolidated sedimentary rocks, for example, when measuring hydraulic conductivity values and when using these values in models predicting groundwater flow or the transport of heat or solutes.
- Properties of rocks controlling groundwater flow if inertial forces or turbulent friction cannot be neglected, particularly how properties of fracture or conduit porosity control groundwater flow dynamics and aquifer properties observed at large scale.
- Feedbacks between groundwater flow, chemical and thermal evolution of groundwater, and biological processes in different types of hydrogeological settings, including resulting changes in aquifer properties.

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## Further reading

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## Relevant websites

https://gw-project.org/—The Groundwater Project—Online Platform for Groundwater Knowledge.
https://www.un-igrac.org/what-groundwater—International Groundwater Resources Assessment Centre—Basics, Glossary, and Stories about groundwater.
https://www.usgs.gov/special-topic/water-science-school/science/groundwater—U.S. Geological Survey—Groundwater Science.
https://iah.org/education/—International Association of Hydrogeologists—Education Resources.