Chapter 3
Characterization of Karst Aquifer

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Proper knowledge of an *aquifer system* is prerequisite for its utilization, protection from pollution, and sustainable development. To recognize and understand a karstic system fully is a very long and difficult process, one which will probably never be completed. Why? Simply the reason is that the system is so complex and heterogeneous. However, during the last century, karst scientists took many important steps forward and today, we know much more about karst system properties than did our predecessors, the founders of karstology.

Could the recognition of a natural system such as karst be reasonably compared with the recognition of a human being? This question may well seem strange, but there is actually some sense behind it. Initially, information is usually collected “second-hand” and from remote sources, and this introductory information could be similar for both: name, address (location), and some general data. Then, the subject is met and its size and shape observed which might be sufficient to create a very general overall impression. Then, the most sensitive step, talking or surveying starts, and word by word, or applied method by applied method, information on personality, i.e., on system behavior, is collected. While the external shape or figure of the human being could be equated to the geometry of the aquifer, the behavior of the karstic aquifer system could be identified with the following properties of aquifer: permeability and storativity, flow pattern and direction, processes of recharge and discharge, natural water quality, and several others. And many of these properties are very variable in space and in time. Therefore, it is a very

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delicate task first to collect all necessary information and even more so to have the *character* of each individual karstic aquifer system properly explained in order that it may be fully understood.

What is the *aquifer system*? First, aquifer is usually defined as a porous media (rock mass) that can store, transmit, and discharge an important amount of water. "Important" here does not necessarily mean "significant" or "huge," but refers rather to an amount sufficient to be observed and economically valuated to satisfy demands of the consumer(s). If groundwater quantity within the rock media is too small, instead of "aquifer," the terms "*aquitard*" and "*aquifuge*" are used. The latter means an almost total absence of water in rocks. In reference to karst, *aquifer* is almost always discussed even though there are many compact carbonate rocks with low permeability. Finally, the attribute "system" indicates the complexity and functionality of an aquifer.

How can karstic aquifers be distinguished? There are several criteria, of which just a few will be emphasized here.

In accordance with *discharge* and accordingly with a transmitted and stored amount of water, there are

- Karst aquifer of high productivity,
- Karst aquifer of moderate productivity,
- Karst aquifer of low-to-moderate productivity (Box 3.1).

When an aquifer produces just a small amount of water (low productive), the terms *fissured* or *fractured* are more appropriate than *karstic* because the system of voids and joints is probably not karstified, i.e., not expanded under the influence of mechanical and corrosional agents. It is also common to entitle aquifer of low-to-moderate productivity as combined *karstic-fissured aquifer*, or vice versa.

Box 3.1

On the Hydrogeological map of Dinaric karst prepared under the GEF regional project Protection and Sustainable Use of the Dinaric karst transboundary aquifer system (DIKTAS), implemented by UNDP and executed by UNESCO, the following karst and fissured aquifers are distinguished (Fig. 3.1):

1. Highly productive karst aquifer (KA1),
2. Moderately productive karst aquifer (KA2),
3. Fissured aquifers (FA).

In total, 15 lithostratigraphic units are grouped to KA1 in accordance with aquifer permeability, transmissivity, flow type, and discharge. Nineteen units are classified into the second group KA2, while seven are attributed to the FA group (Stevanović 2011).

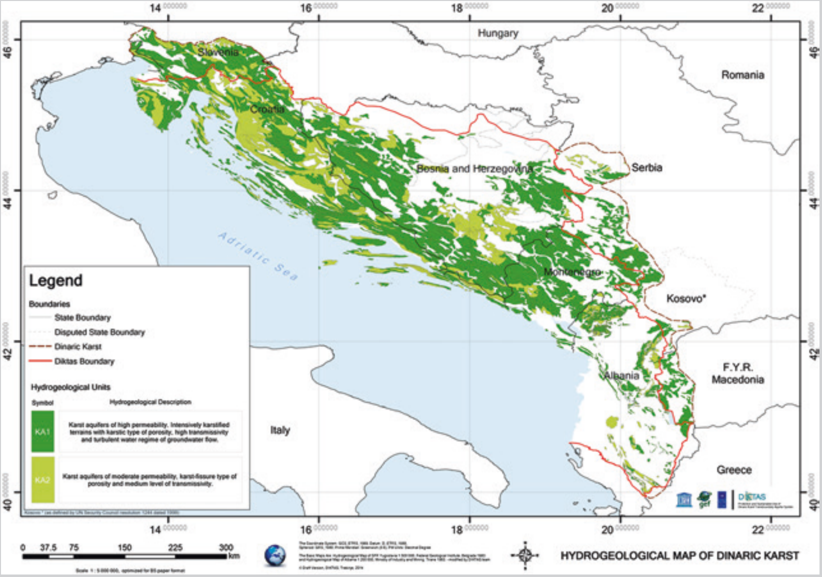


Fig. 3.1 Extension of highly productive (KA1) and moderately productive karst aquifers (KA2) in Dinaric karst (http://dinaric.iwlearn.org/, printed with permission)

According to the dominant type of rocks of which karstic aquifer consists, a classification similar to that presented in Sect. 2.5 can be made as follows:

- Carbonate karst aquifer,
- Dolomitic karst aquifer,
- Marble karst aquifer,
- Chalky karst aquifer,
- Anhydritic karst aquifer,
- Gypsum karst aquifer,
- Halitic karst aquifer.

Bakalowicz (2005) separated carbonate aquifers into the following major groups: the fracture, non-karstic aquifers, and the truly karst aquifers, with all intermediate stages of karst conduit development.

Finally, taking into consideration structures and hydrodynamic properties, there are

- Unconfined karstic aquifer,
- Confined karstic aquifer,
- Semi-confined karstic aquifer.

The first group is characterized by the absence of any cover, free *water table,* or *potentiometric pressure* in karstic voids equal to atmospheric at concerned depth. Water table as surface-connecting water table in voids represents the upper limit of saturation (the term *phreatic* is also commonly used, Fig. 3.2a). *Confined* karst aquifer lies below the impermeable cover, and its *potentiometric pressure* or *hydraulic head* are sufficient to raise the groundwater level over the base of the overlying

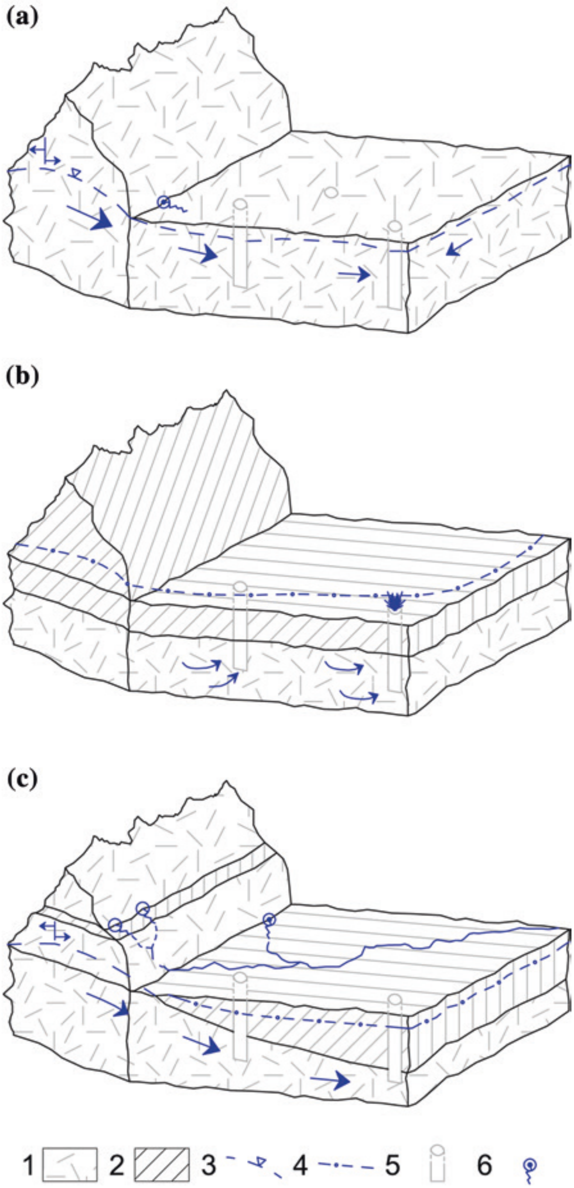


Fig. 3.2 **a** Unconfined karstic aquifer. **b** Confined karstic aquifer. **c** Semi-confined karstic aquifer, and perched aquifer within its upper part drains through small springs. *Legend 1* karstic aquifer, *2* impervious rocks, *3* groundwater table, *4* hydraulic head (piezometry), *5* borehole, and *6* spring

bed when it is penetrated by the borehole (Fig. 3.2b). *Semi-confined* (transient) karst aquifer includes confined and unconfined sections. *Perched aquifer* is the term usually applied to an aquifer lens localized within aquitard ("pockets"), or isolated unconfined beds of limited extension above the regional water table. In the latter case, perched aquifer is disconnected from deeper parts by some impervious layers or by non-karstified rocks (Fig. 3.2c). Younger (2007) stated that great care must be taken not to confuse perched aquifers with regional unconfined aquifer below.

3.1 Aquifer Geometry and Elements

The size of the aquifer is defined by its boundaries. They are *lateral or horizontal*, and *vertical*, while in accordance with watertightness, the boundaries are *permeable, semipermeable,* and *impermeable*. In the last case, there is a factual aquifer limit, while if boundaries are permeable or even semipermeable, there is just a lithological break, but no water barrier because circulation through the boundary is possible (*relative barrier*). The term *groundwater body,* widely introduced by the Water Framework Directive of the European Union (EU WFD 2000), considers the case of possible linkage of two or more vertically or laterally interconnected aquifers.

The subsurface catchment in most cases is different from the topographical (Fig. 3.3). Herak (1981) stated that the actual catchment area of the Cetina River in the Croatian part of Dinaric karst is 2.7 times greater than its topographical frame. This case and several other cases are discussed in Bonacci's reference book on karst hydrology, *Karst hydrology with special reference to the Dinaric karst* (1987) and are presented by the same author also in Chap. 5.

The flexibility and variability of *lateral boundaries* of a karst aquifer is also discussed in Chap. 6. It can be the result of variation in the water table during the high- and low-water periods or of a possible temporary reorientation of flow direction. Herak (1981) concluded that most of the catchment areas are asymmetrical and only approximately determinable even in regions where intensive survey has been conducted.

Concerning the determination of *vertical boundary,* the case of the top boundary is much simpler than the bottom one. The top is defined by land surface or

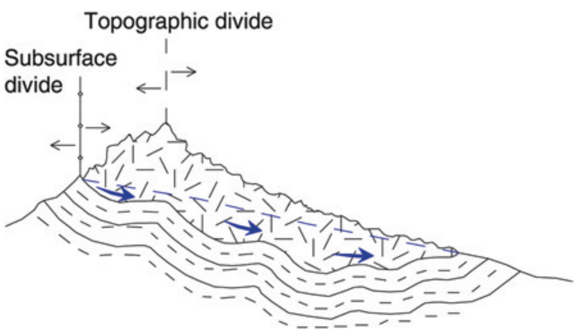


Fig. 3.3 One of the very typical and not rare cases of significant difference between topographical and underground catchment

in the case of confined aquifer by the base of the overlying bed. The bottom boundary is either a *karstification base* or the top of an underlying bed. According to Milanović (1981, 1984), a karstification base is a boundary which is not well-defined and can be considered a transitional zone between karstified and non-karstified rocks.

To locate an even approximate karstification base can be very difficult, but it is an important task of hydrogeological survey. As more information from drilled cores, down-the-hole camera recording and geophysical logging is collected, the chance for relatively correct approximation of this imaginary line increases.

The field hydrogeological survey and tracing tests are essential methods for assessment of karst surface geometry and delineation of the catchment. This topic is discussed in Chap. 4. In hydrogeological practice, it is also common to apply an inverse method to estimate the recharge area based on the results obtained from water budget calculation. Although uncertain in many respects, the method may nonetheless still provide a general view of the geometry of the studied aquifer, especially if similar karstic terrain and aquifer are properly explored in terms of permeability and storativity whereby the analogy between the two is somehow confirmed (Box 3.2).

Box 3.2

Petrič (2002) calculated the recharge area of the Vipava springs (Slovenia) based on observed rainfall data, estimated evapotranspiration rate from the relevant maps and average discharge of the springs covering the 30-year period from 1961–1990. From the average rainfall sum of 2,075 mm/year, and an evapotranspiration rate of 640 mm/year, the difference of 1,435 mm/year has been obtained. This value, which can be equalized with effective infiltration to karst aquifer due to minimal runoff, was compared with the average spring discharge of 6.78 m^3/s. Although there was almost no difference between the two values, Petrič found that the total catchment area is 149 km^2.

$$F = Q/I_{ef}$$
$$F = 6.78 \times 31.536 \times 10^6\,m^3/1.435\,m$$
$$F = 149\,km^2,$$

where

F catchment area of the springs,
Q total annual spring discharge,
I_{ef} effective infiltration on annual basis.

From the geological map, it is concluded that flysch impermeable rocks cover some 9 km^2 (allogenic recharge area), while carbonate karst aquifer covers the remaining 140 km^2.

The method of multiple nonlinear correlations applied to the karst spring hydrograph for the assessment of the catchment area is discussed in Sect. 15.2.

The elements of a typical karst aquifer are the following:

- Top surface,
- Epikarst,
- Vadose zone,
- Saturated zone, and
- Karstification base.

When might the three first elements and the last one be absent? The answer is in the case of confined karstic aquifer. It is thus only the presence of a *saturated zone* that is certain, but if saturation is also missing, then rocks are dry and in fact, we are not dealing with an aquifer.

The schematic representation of an unconfined karstic aquifer and its elements is shown on Fig. 3.4.

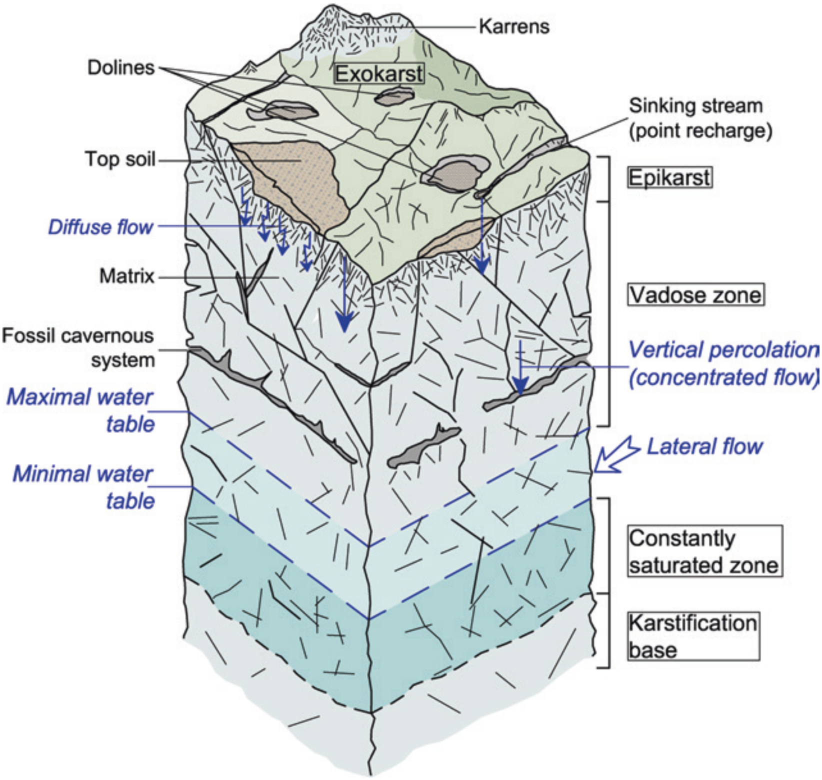


Fig. 3.4 Model of unconfined karst aquifer

The *top surface* or *exokarst* may contain variable features, some of which stimulate aquifer recharge, while others obstruct this process. For instance, dolines could have a dual function: As a depression in relief, they collect water which circulates to their bottoms, but if the bottom is filled with thick and impermeable deluvial soil, the water will remain in the soil or just refill the formed swamp until it finally evaporates (Fig. 3.5).

The top surface includes soil horizons resulting from the weathering process, but soil can also be brought about artificially. It is common in soil-poor karst terrains for local villagers from generation to generation to destroy rocks and expand soil surface and the thickness of cultivated land (Fig. 3.5).

An unsaturated zone usually consists of two parts: an upper zone or epikarst and an underlying vadose zone.

Epikarst is a "skin" of the karst system underlying the soil cover and opens to the surface (Figs. 3.6 and 3.7). Its role in the functioning of a karst system,

Fig. 3.5 Doline—swamp in high karstic plateau of Durmitor Mt. (Monetnegro) (*left*) and cultivated bottom of doline in Vrachanski karst (NW Bulgaria)

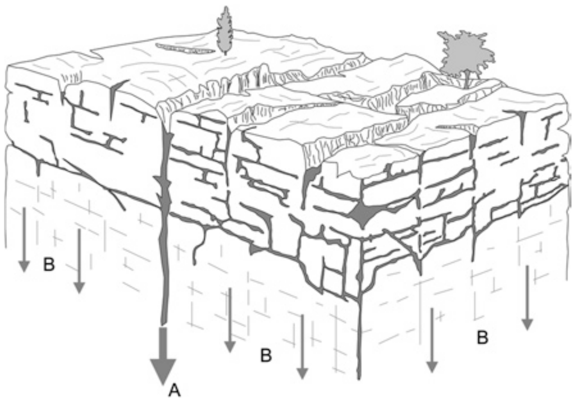


Fig. 3.6 Scheme of the epikarstic aquifer in the zone of alteration and superficial fracturation. *A* is vertical drainage through large fissure, *B* is seepage, i.e., effluent phenomenon (from Mangin 1975)

Fig. 3.7 Epikarst covered by thin soil layer (*left*) and small discharge spring draining perched aquifer (*right*)

in particular its functions of recharge and discharge, has been studied by many authors. Mangin (1974, 1975) explained the functions of a karstic system locating "epikarstic aquifer" above a real "karst aquifer" and also suggesting the term *eukarstic aquifer*. Atkinson (1977) was among the first who explained the epikarst drainage mechanism and its impact on the entire spring hydrograph. Williams (1983, 2008) called epikarst a *subcutaneous zone* and analyzed its role in karst hydrology and cave hydrogeology. He emphasized that porosity and permeability are in principle greater near the surface than down deep.

Kiraly et al. (1995) attributed the base of epikarst to *Faraday cage* with respect to lower aquifer parts. The hydrodynamic analyses on a 3-dimensional model of finite elements indicated that epikarst has a large impact on (1) The shape of karst hydrograph, (2) The base flow component, (3) The water table fluctuation in conduit network and surrounding rock blocks, and (4) The recharge of low permeable rocks.

Klimchouk (2000) highlighted the diffuse karstification of the underlying bulk rock mass and the higher degree of fissuring of the epikarstic zone. He stated that contrast in the hydraulic conductivity of these two zones allows some groundwater storage in the epikarstic zone and flow concentration in its base. Trček (2003) further discussed diffuse vertical recharge from epikarst which decreases with depth and converges into lateral flow. Perrin et al. (2003) analyzed the storage in epikarst based on isotopic data and developed a conceptual model. Bakalowicz (2005) emphasized production of carbon dioxide in a perched saturated zone. This CO_2 as an important karstification agent is then transported to deeper aquifer parts.

According to Ford and Williams (2007), the epikarst is typically 3–10 m deep. It consists usually of a particularly weathered zone of limestone which gradually gives way to the main body of the vadose zone that comprises largely unweathered bedrock.

Common among the explanations provided is the existence of perched aquifer in the uppermost part of the vadose zone resulting in water retention and a delay in water infiltration to the lower parts. This delay depends on the epikarst permeability and may take a few days or a few months.

Can epikarst be absent? Krešić and Mikszewski (2013) aptly stated that the concept of epikarst was applied indiscriminately to explain aquifer behavior and functioning. They stated that epikarst is mentioned in many studies as a factor influencing aquifer regimes even for terrains where it is completely absent and no perched aquifer is present (Figs. 3.8 and 3.9).

We may conclude that epikarst, when present, is located within the uppermost part of the vadose zone and is partly saturated, storing a certain amount of water and rerouting vertical infiltration to the deeper phreatic zone of the karst aquifer.

Fig. 3.8 Lack of epikarst. Almost unique degree of fissuration in vertical section in massive limestones of Triglav (Slovenia, *left*) and Verdone (France, *right*)

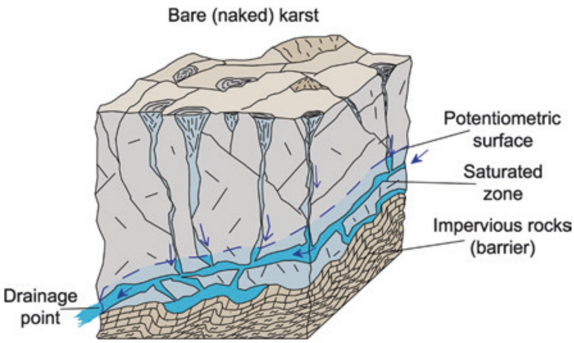


Fig. 3.9 A scheme of terrain with dominant bare (naked) karst. Very deep potholes and cavities very quickly convey water to the saturated bottom part. This situation is typical for Dinaric holokarst

In contrast, in many highly karstified terrains water from rainfall or sinking streams infiltrates so quickly to the saturated zone that we have completely dried caves and fossil channels and a total absence of water in the uppermost part, as discussed in Chap. 2. Therefore, the field hydrogeology survey and the assessment of the presence of perched aquifer and epikarst and its thickness and permeability must precede any hydrodynamic analysis or modeling of karst aquifer regime.

The vadose zone is a deeper, transient zone in karst aquifer. In the vertical direction, it ends with the groundwater table. The terms *unsaturated zone* or *zone of aeration* are also commonly used. The latter refers to the atmospheric pressure and to the fact that some of the voids may contain air. The circulation in the vadose zone is dominantly vertical, but along with the transfer of water which percolates from the top surface or epikarst to the water table, some lateral flows also may occur. This strongly depends on the orientation and angle of karst voids and channels. Therefore, preferential paths for water circulation are created in the matrix, fissures, and caverns. Parizek (1976) presented a strong relationship between the main water paths and large fault zones (Fig. 3.10).

The saturated zone represents the main aquifer layer and water storage space. Its thickness depends on several factors such as recharge, permeability and storativity, position and size of discharge points, hydraulic head, strike and dip of bedding planes, and location and utilization of artificially introduced intakes. Water also flows through the system of interconnected voids, joints, fissures, and cavities of various sizes.

Figure 3.11 is a visual interpretation after Drogue (1982) of the system of different elements for water circulation. The various shape and size voids and cavities are separated into blocks. Their density is in accordance with the vertical position: The maximal density but small fissures is in the first level blocks, closest to the surface. There is now an essential difference in the size of voids and cavities which belong to the blocks in the second and third deepest level, but the density decreases with the depth. However, the main drainage system at the deepest part can be enlarged by mechanical erosion of lateral flows which are concentrated near discharge points.

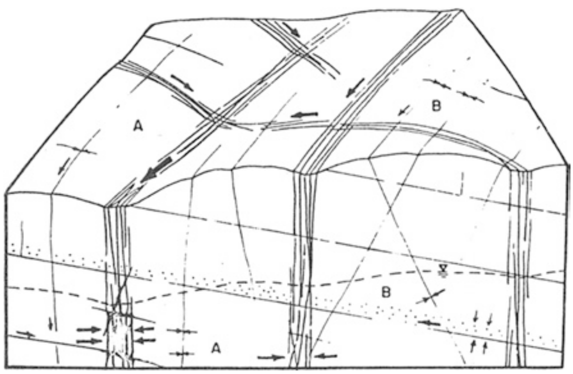


Fig. 3.10 Massive blocks in carbonate rocks dissected by large faults and fissures which enable water infiltration and circulation (from Parizek 1976)

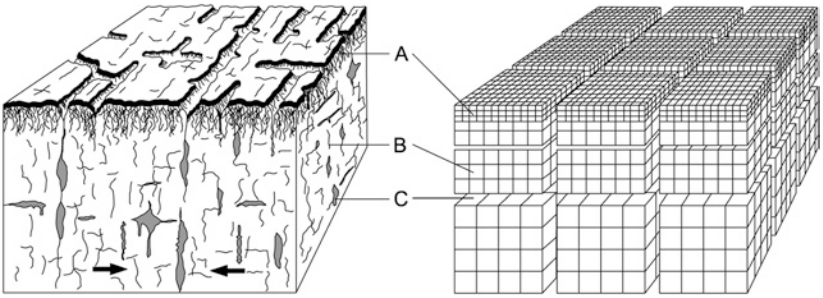


Fig. 3.11 Scheme of karstic blocks. *A* The uppermost blocks are altered, very permeable and allow rainfall infiltration. *B* Mid blocks, less permeable with fissures and small cavities as the dominant elements. *C* Deepest block, also fissures and cavities, and locally larger caverns which evacuate (drain) water from the system (modified from Drogue 1982)

As mentioned already the *karstification base* is an aquifer bottom, relatively located and non-fixed. The depth to the *karstification base* can be very large, sometime thousands of meters (Box 3.3). In the Dinaric karst, the exploratory borehole found cavities in Permian–Triassic clastic limestones at a depth of 2,236 m (Milanović 1981). However, local karst development should always be differentiated: For example, small paleo-cavitation which is disconnected from the rest of the voids in a rock mass should always be differentiated from the real karstification which results in interconnected cavities that enable active circulation (*effective porosity*).

Box 3.3

- Milanović (2006) evaluated results of pressure permeability tests in 140 boreholes drilled in eastern Herzegovina. The karstification exponentially decreased, and it is almost 30 times greater in the subsurface zone than at a depth of 300 m. He also stated that at greater depths, karstification is developed along major faults.
- The case of inverse karstification, greater at bigger depths and developed along restrictive pathways, is described by Stevanović (2010a). In the Bogovina area in the Carpathian karst of eastern Serbia, the local maximal karstification depth has been estimated to be at over 800 m, while karstified intervals with elements of paleo-karstification have been confirmed 471 m deep on a borehole SB-1. However, the major karstified zone was identified during the drilling at a depth of 60–90 m. This is a very active conduit system, which conveys most of the local groundwater. Some of the large caverns in that system are filled with secondary deposited clays and sands and are not currently active.
- Very similar variable porosity (cavernosity) with a maximum depth between 65 and 125 m has been recorded during the drilling of several exploratory boreholes in karst of Zagros Mt. in Iraq (Fig. 3.12).

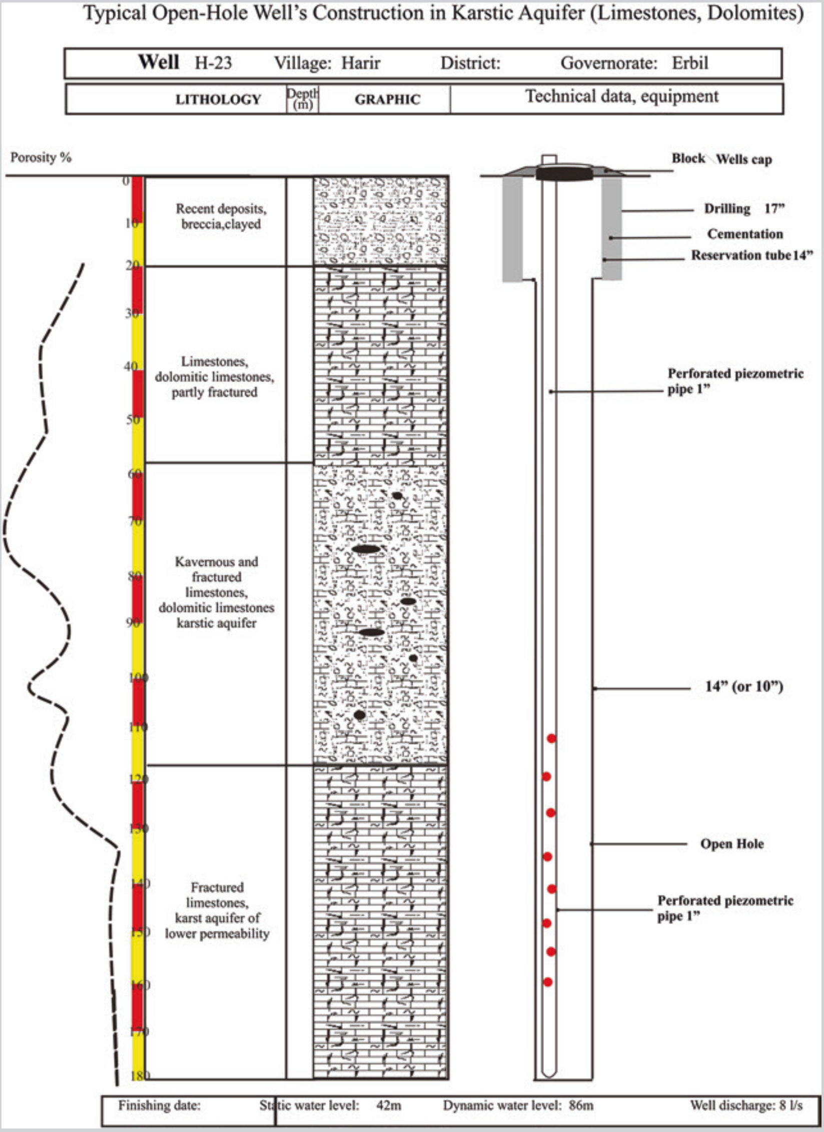


Fig. 3.12 Drilling log of the borehole drilled in the Harir area (Erbil Governorate, Iraq). Porosity curve indicates maximum cavernosity at the three intervals: 70, 100, 120 m

3.2 Permeability and Storativity

In Chap. 2, the type of rocks subjected to karstification was noted. In addition, it was highlighted that not all rocks are equally soluble and that mechanical and chemical agents are crucial for the karstification process as is the presence of water and pores. The *porosity* refers to the proportion of volume of rock occupied by pores. The two types of porosity exist: *primary* and *secondary*. The former is the result of rock diagenesis, while the latter results from tectonic fabrics, exogenetic factors, and karstification. Theoretically, the final result of the karstification process is the total conversion of primary into secondary porosity.

In principle, the matrix of the majority of karstic rocks has a small *primary porosity* commonly referred to as a *microscopic porosity*. This particularly concerns limestones and dolomites (Fig. 3.13). Castany (1984) noticed that chalk consisting of calcite particles has *interstitial porosity*. Synonymous with that term is the term *vuggy porosity* which relates to visible pores between some of the constituent components (Fig. 3.14).

The *secondary porosity* is also called *macroscopic porosity* which is a visible porosity that could be presented in poorly cemented or oolitic carbonates but which mainly refers to bedding planes, joints, fissures, fractures, and cavities formed by a dissolutional, and not diagenetic process (Fig. 3.15). The secondary

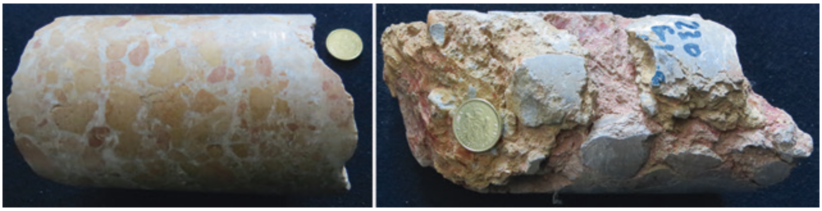


Fig. 3.13 Carbonate breccia: well cemented and compact with very small porosity (*left*) and poorly cemented (*right*)

Fig. 3.14 Vuggy porosity of Miocene marine limestones, the widely spread formation in the Alpine karst (named Asmari Fm. in Iran, Pila Spi Fm. in Iraq, Sarmathian Fm. in Balkan countries)

Fig. 3.15 Secondary porosity, but restricted: cavities partly filled with recrystallized calcite in limestone core samples

porosity is responsible for transferring gravitational water, but if voids are not connected, there would still be no free flow system created in the aquifer. For this reason, hydrogeologists must try to distinguish the *total porosity* from *effective porosity* (Box 3.4).

The total porosity is thus the ratio between the total volume of voids and the total volume of rock (sample, core). Voids are, therefore, not necessarily connected and enable water circulation. According to Castany (1984):

$$P = V_v / V \tag{3.1}$$

where

P total porosity,
V_v volume of voids,
V total volume of rock.

In contrast, the *effective porosity* is the volume of interconnected voids against the total volume of rock.

$$P_e = V_e / V \tag{3.2}$$

where

P_e effective porosity,
V_e volume of interconnected voids,
V total volume of rock.

Some authors also recognize *tertiary porosity* which creates very large caverns. In accordance with explained differences, we may distinguish the following:

- matrix porosity, as mainly primary,
- fissure (fracture) porosity, as secondary, and
- cavernous porosity, as tertiary.

The presence of two or all types of porosity is termed *dual porosity*.

Box 3.4

Castany (1984) presented total porosity of different kinds of carbonate rocks taken from several locations. The values range widely from only 0.1 % for Carrara marbles in Italy to 45 % for chalk in France. Ford and Williams (2007) also found that matrix porosity of some samples of chalk from England is around 30 %, but its secondary porosity is very low, less than 0.1 %.

Krešić and Mikszewski (2013) presented the case of Biscayne aquifer from Florida, with very high primary porosity of 40–50 % which is increased further by karstification.

Milanović (1981) calculated effective porosity of littoral karstic aquifer in carbonate rocks to be in the range of 1.4–3.5 %, based on results of analyses of recession discharge curve (see Chap. 7 for method explanation) of the Ombla Spring near Dubrovnik and groundwater fluctuations. Milanović also cited data of Vlahović (1975) and Torbarov (1976) who estimated effective porosity in neighboring Dinaric karst aquifers in Nikšić karst polje and in the Trebišnjica springs basin. Values of around 6 % for the highly tectonized zone in Nikšić and 1.2–1.5 % for the entire Trebišnjica basin are obtained.

Secondary porosity does not mean that fissures and cavities, even interconnected ones, always remain open for water circulation. The karstic paleo-flows, and especially mineralized and thermal waters, have caused many previously active systems to become filled with sediments such as clay, sand, gravel, or calcite and aragonite (Fig. 3.15). If the plug is compact, solid and recrystallized, then permeability in this segment is lost, but if unconsolidated sediments are present, then long and forced pumping or big floods which pressurize water flow may clean out cavities and reactivate them. This explains why water turbidity at karstic outlets (springs) may significantly increase, and this is also the reason why hydrogeological parameters which define karst aquifer behavior change over time.

Effective porosity is often equated with *specific yield*. Castany (1984) highlighted their differences: Effective porosity is an index of available interconnected pore or cavity space, whereas specific yield is the ratio of the amount of water which can freely flow by gravity from the rock against the total volume of the rock. Younger (2007) defined specific yield as the "amount of water which drains freely from the unit of volume of initially saturated rock per unit decline in water table elevation." The point is that not all water will flow out from the pores or cavities, and specific yield is thus smaller than effective porosity for *specific retention*.

Analyzing the effective porosity and infiltration/filtration capacities of rocks, Younger (2007) introduced the terms "fillable effective porosity" and "drainable effective porosity." Considering that some adhered water always remains in an aquifer and that “new” water can only fill dry pores, voids, or cavities, he declared *fillable effective porosity* to be less than effective porosity. *Drainable effective porosity* is something similar to the above described specific yield: effective porosity minus specific retention.

The two main characteristics of karstic aquifers are *anisotropy* and *heterogeneity*. *Anisotropy* means that one physical property varies with direction. In the case of a karstic environment, anisotropy is very typical and has almost become the first association with the word “karst.” A great variation in fissuration or cavernosity in the vertical section has already been confirmed, but at the horizontal level compact blocks, small fissures and also large cavities may be jointly present. The latter refers more to *heterogeneity* as a variation of a property from site to site within the same formation (Box 3.5). While in an unkarstified rock heterogeneity within a common permeability area may be perhaps 1–50, in karstic rocks it may increase to perhaps 1–1 million (Ford and Williams 2007).

Box 3.5

Figures 3.16 and 3.17 show various properties of karstic and karstic-fissured aquifers recorded during the intensive field survey and tests conducted for the regional water supply system of Bogovina in the Carpathian karst of eastern Serbia (Stevanović 2010b).

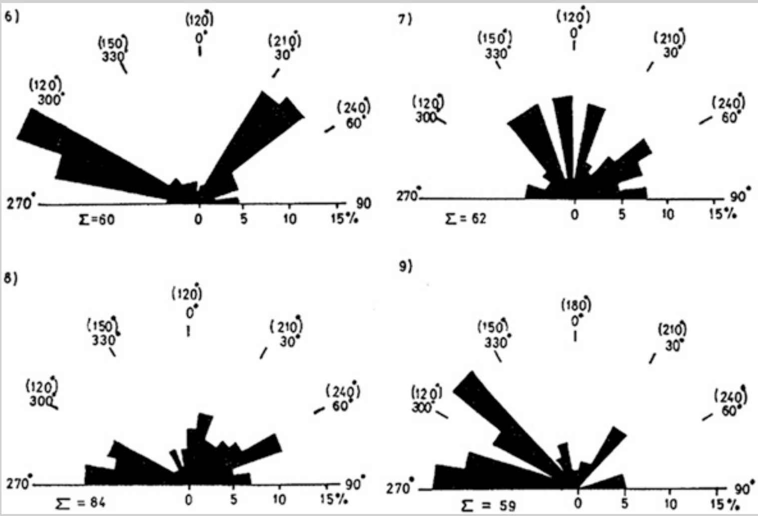


Fig. 3.16 Rose-chart diagrams of the orientation and frequency of measured fissures and cavities within four selected blocks in Aptian limestones

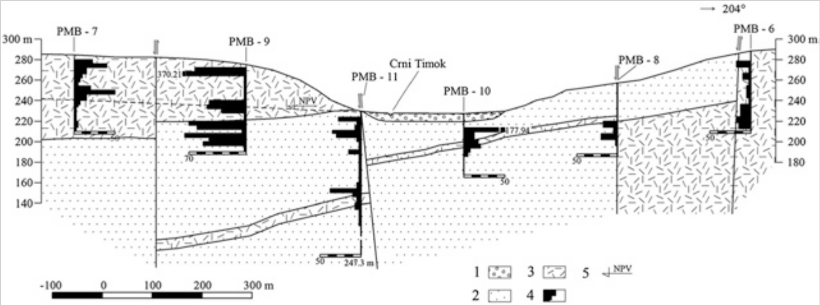


Fig. 3.17 Results of water injection tests conducted in Aptian limestones and sandstones at Bogovina dam site show considerable differences in permeability within the same blocks. *Legend 1* alluvium, *2* sandstones and sandy limestones, *3* karst, *4* volume of water injected per intervals, and *5* water table

Krešić (2013) noticed that the density, size, and geometry of individual conduits in the karst aquifer vary enormously depending on specific properties such as mineral composition, tectonic fabric, recharge mechanism, and position of erosional base.

Such differences are clearly evidenced in the case of seven piezometers situated very close together and located at short distances from and perpendicular to the Trebišnjica River (Bosnia and Herzegovina). This case is described by Milanović (2006). During the summer months, the riverbed is dry, while the groundwater table is below the bottoms of all piezometers. During the floods and river flow of 10–30 m^3/s, the groundwater table rises but not to over a depth of 23 m. Although in the same lithological unit, and in close proximity, the amplitude of the differences in water table of those piezometers is 17 m. This is the result of the permeability of the overlying sinking zone (ponors in riverbed) and aquifer heterogeneity (Fig. 3.18).

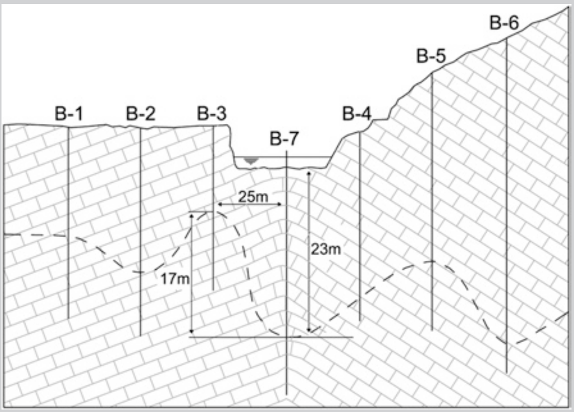


Fig. 3.18 Cross section of karst aquifer in the Trebišnjica River basin, Kočela locality. The position of the water table of an unconfined aquifer in a high water period is marked with a broken line (from Milanović 2006, printed with permission)

The last presented case confirms what hydrogeologists often realize in daily practice: A distance of just 1 m or even less is enough for completely different drilling results: One borehole in a completely compact limestone block without a single drop of water and another which attains a privileged underground water course with large discharge and small drawdown (see Box 3.9).

The *anisotropy* and *heterogeneity* influence other aquifer properties—*permeability* and *hydraulic conductivity*, *transmissivity,* and *storage coefficient* as standard hydrogeological parameters. The recommended references for these topics are those written by Meinzer (1923a), Castany (1984), Palmer et al. (1999), Kiraly (2002), Ford and Williams (2007), Worthington and Ford (2009), Krešić (2013). We shall refer just briefly to the terminology and provide short explanations necessary to understand some further topics.

Permeability is a property of aquifer that allows a fluid to flow under a pressure gradient. It can also be defined as the ability to convey water through the rocks. This aquifer property is quite different from effective porosity, a fact which surprises many non-hydrogeologists. Although effective porosity indicates the proportion of rock sample or mass occupied by interconnected pores, for permeability it is the size of the openings which connect pores and voids that is important. Therefore, limestones with a very small porosity, e.g., 1 % or even less, may be very permeable because their connected cavities as a preferential path may transmit a large amount of water. In contrast, chalk with a high effective porosity is much less permeable: Due to small pores, specific retention of chalk is high. Younger (2007) stated that the list of rocks with both high effective porosity and high permeability is very long, but in any case, effective porosity is the prerequisite for permeability.

The difference in *hydraulic head* and/or *potentiometric pressure* is a driving force for water flow. *Head* is the result of measured water pressure and atmospheric pressure at a given point. *Potentiometric pressure* is a theoretical and imaginary surface connecting all points to which water released from aquifer (confined) would rise. If this rise enables a free flow from the well, such a well is called an *artesian well*. The same term is applied to the potentiometric pressure at that exact point: *artesian pressure* (the term derives from an old self-flowing well in Artois, Belgium, but today is rarely used because a free flow from a well can also result from other factors).

The difference in potentiometric pressure to the distance over which the change occurs is called the *hydraulic gradient*. Its presence is prerequisite for water movement, but velocity still very much depends on the permeability and effective porosity of the aquifer (Box 3.6). If the size of openings is too small, for instance of a microscopic scale, water will move very slowly. In karst that could be the case with sediments which fill the conduits and in that case molecular diffusion could be more active than the impulse of the natural hydraulic gradient (Younger 2007).

Box 3.6

In the case of karst and high pressure (hydraulic gradient), the water can be propagated very quickly to great distances. Many tracing tests conducted in karst worldwide confirmed velocities of a few centimeters per second, and even up to tens of centimeters per second, which may be approximated to the existence of a waterfall in the karst ground.

Referring to 380 conducted experiments in Dinaric karst, Komatina (1983) concluded that the frequency of fictive groundwater velocities in Dinaric karst is as follows: in 70 % of cases from 0 to 5 cm/s; in 20 % of cases 5–10 cm/s; and in 10 % of cases more than 10 cm/s. Groundwater velocity directly depends on the hydrological period and the water table position (Fig. 3.19).

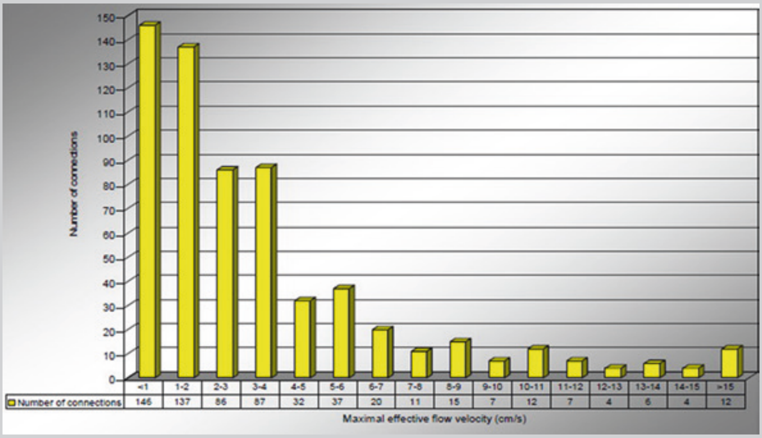


Fig. 3.19 Frequency of maximal flow velocities based on 623 confirmed ponor spring connections in Croatia (after Pekaš, DIKTAS database http://dinaric.iwlearn.org/)

During the dry season and low aquifer water table, water circulation in the karst system is characterized by a slow movement of aquifer waters. But the water waves labeled with dye take two- to five-fold less time to travel the same distance during a season of high hydrological activity (Milanović 2001). For instance, to cover the distance (34 km) from Gatačko Polje to the Trebišnjica Spring (Dinaric karst, Herzegovina), the underground flow takes 35 days when the water table is low and inflow is small. During the high water levels and large inflow, the well-distinguished water wave takes only 5 days to cover the same distance (Milanović 1981). Similarly, between Čaprazlije ponor and Mali Rumin Spring, also in Dinaric karst, the velocity was as follows: in the dry period 5.1 cm/s; at high water level 28.8 cm/s (Komatina 1983).

A large velocity value has been obtained by tracing tests conducted in Prespa Lake, connected with the Ohrid Lake (Macedonia/Albania, see details in Chap. 5). The maximum values in the test conducted in 2002 were between 19 and 80 cm/s.

In principle, Darcy law (1856) expressed as a quantity of water flowing through homogenous intergranular aquifer which is proportional to the difference in pressure (hydraulic gradient) is not applicable in karst. The reasons have already been mentioned: *anisotropy* and *heterogeneity*. An adequate and detailed explanation can be found in Krešić's works (2007, 2013), but perhaps the simplest way to understand why not to use Darcy law is provided by Krešić and Mikszewski (2013): The hydraulic head may go up and down along the same pipe as the cross-sectional area increases or decreases, respectively.

However, in some cases, an approximation of homogenous media can be applied with caution. Such exceptions include, for instance, the prevalence of matrix porosity, or a fissure–karstic system of limited permeability which allows only laminar, but not turbulent flows, or chalk with vuggy porosity with prevailing laminar and diffuse flow. In such limited circumstances, an approach called equivalent porous medium (EPM) could still be acceptable. But many mathematical EPM models were developed based on the inclusion of "problematic" privileged paths (cavities) and of course there have been many negative results when the models have been applied.

If *Darcy law* is not viable in karst, how then can parameters which define hydraulic properties of the aquifer be managed? *Hydraulic conductivity* (*K*) is probably the most problematic because it directly implies homogeneity (conductance) of intergranular porous media. Hydraulic conductivity (*K*) is equal to groundwater flow (*Q*) divided by the hydraulic gradient (*i*) and cross-sectional area of flow (*A*):

$$K = Q/i \cdot A \tag{3.3}$$

$$i = \Delta H/L \tag{3.4}$$

where

ΔH difference in hydraulic head between two observation points (loss of energy), and
L distance between observation points.

The unit for *K* is the same as the unit for velocity (m/s). The feasible application considers knowledge of conductivity in all three dimensions or directions (K_x, K_y, K_z), but this would never be an easy task in karst simply because all elements of the karstic system, such as the position of bedding planes or openings and the orientation of every fracture (dip and strike), will never be known, no matter how detailed a survey is conducted. Figures 3.20 and 3.21 show some examples of great differentiation in the development of privileged groundwater paths identified during a hydrogeological survey in northern Algeria.

Another parameter *transmissivity* (*T*) (or *transmissibility*) integrated hydraulic conductivity (*K*) and saturated thickness (*h*):

$$T = K \times h \quad (\mathrm{m}^2/\mathrm{s}) \tag{3.5}$$

The aquifer thus produces more water (is more transmissive) when hydraulic conductivity and saturated thickness are greater. In confined aquifer, thickness is stable except when over time a high pumping rate causes a gradual reduction

Fig. 3.20 Subhorizontal karstic channel in totally impermeable block of Upper Cretaceous limestones. It functions as a single conductive path for groundwater, but it is able to transfer enormous flow (foundation of the Hammam Grouz dam, Algeria)

Fig. 3.21 Core samples taken from the exploratory borehole at Ourkiss dam site (Oum el Bouaghi, Algeria). Great variation in lithology of Turonian limestones: fissuration degree, texture, and impurity. No sample taken from a section of around 1 m in length indicates the presence of a cavern

of pressure and the aquifers become unconfined. In contrast, the saturated thickness in unconfined aquifers is variable and depends on fluctuation of water table. Therefore, transmissivity is variable, too. Knowing that fast changes in the water table in karst are very common, it can be concluded that not only anisotropy is an obstacle to the use of transmissivity as an aquifer property, but also too is the regime of karst aquifer. However, considering that most laboratory and infield tests result in a calculation of K and T and that there are just a few other options for the assessment of hydraulic behavior, application of these parameters in hydrogeological practice is still inevitable (Boxes 3.7 and 3.8). This application must, though, always be done with caution and with comparison of the results obtained from one site to another.

The effective storage volume in karstic aquifer depends on its area, aquifer thickness, effective porosity, maximal water level, and karstification base (Issar 1984). *The storage coefficient* (S, ε) is equal to *specific yield* in unconfined aquifers. The specific yield is the change in the amount of water in storage per unit area of unconfined aquifer that occurs in response to a unit change in head (Castany 1984). Younger (2007) highlighted that storativity in confined aquifer is a very different physical parameter from specific yield although for both, the symbol S is applied. Ford and Williams (2007) and Krešić (2013) stated that in confined aquifer, *specific storage* (S_s) represents the stored or released volume of water by the unit volume of aquifer per unit surface, due to the change of hydraulic head. The unit is m^{-1} in length. *Storability* or *storage coefficient* for confined aquifer is the product of specific storage (S_s) and aquifer thickness (h):

$$S = S_s \times h \tag{3.6}$$

S is a non-dimensional number and is often presented in the form of % of the total rock mass.

Box 3.7

Figure 3.22 shows a hydrodynamic scheme where the application of standard procedures for calculating hydrogeological parameters supposedly will not result in significant errors and can be used just as an approximation or as "equivalent" parameters as indicated by Marsaud (1997). On the other hand, the pumping test remains the most convenient method to assess aquifer potential for groundwater exploitation.

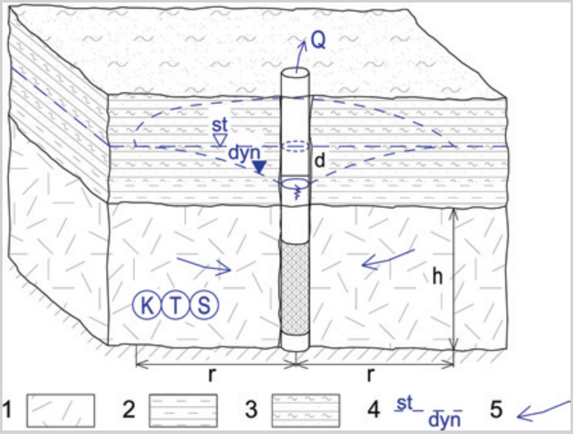


Fig. 3.22 Schematic presentation of a confined karstic aquifer with dominant fissure porosity and presence of small cavities. A fully penetrated well pumps a constant amount of water, and drawdown indicates a non-steady-state flow with slow decline over time. *Legend 1* karstic aquifer, *2* low permeable sand and clay, *3* impervious marls, *4* potentiometric surface, *st* static, *dyn* dynamic, *5* groundwater streamline. Symbols are explained further on in text

Theis (1935) formulated an equation for non-steady-state hydrodynamic conditions which calculates drawdown (d) as a result of pumping rate (Q), transmissivity (T), and pumping well function $W(u)$. Non-steady-state flow means that pressure in confined aquifer declines with time under constant pumping rate.

$$d = (Q/4\pi \times T) \times W(u) \tag{3.7}$$

where

d drawdown in m
Q constant pumping rate in m^3/s
$W(u)$ well function, values $W(u) = f(u)$ including theoretical curve can be found in hydrogeology literature

When $u < 0.05$, the well function $W(u)$ can be expressed with error less than 3 % as follows:

$$W(u) = \ln(2.25 \times T \times t)/r^2 \times S \tag{3.8}$$

where

t time since the beginning of pumping
r radius of well or distance to the point where drawdown is recorded (parameters T, S, and K are associated to sector defined by r)
S storage coefficient

Transmissivity (T) is calculated using Eq. (3.7) and values for d and $W(u)$ which are obtained as matched cross-points of superimposed theoretical curve $W(u) = f(1/u)$, and curve from log-log graph $d = f(t)$ plotted based on results of pumping test:

$$T = Q/(4\pi \times d) \times W(u) \quad (\mathrm{m}^2/\mathrm{s}) \tag{3.9}$$

The storage coefficient (S) is calculated from transformed Eq. (3.8), whereas values for t and $1/u$ are obtained from matched cross-point of superimposed theoretical curve and pumping test curve $d = f(t)$.

$$S = 4T \times t \times u/r^2 \tag{3.10}$$

Finally, hydraulic conductivity (K) is calculated from transmissivity (T) and average thickness of confined karst aquifer (h):

$$K = T/h \quad (\mathrm{m/s}) \tag{3.11}$$

Mijatović (1968) applies Maillet's method (1905) for analyzing spring hydrograph in recession periods by calculating discharge coefficient and volume of discharged waters from aquifer sections which are characterized by different fissuration and permeability. The storativity (S) could be estimated in accordance with the volume of discharged waters and decline of the water table. He also applies this equation for calculating S:

$$S = 2.25 \times T \times t/r^2 \tag{3.12}$$

t time for emptying water reserves from studied aquifer section; other symbols are as above

Storativity (S) could be estimated very roughly by using formula (3.2) for effective porosity (P_e) and calculating the relation of volume of discharged water for the certain period of time (V_t), and the total volume of the rock (V).

$$S = V_t/V \times 100\,\% \tag{3.13}$$

Case example:

The karstic spring discharge has declined from 43.5 to 40 l/s during 1 day. The water table, which has been measured in the well located 300 m from the spring inside the catchment area, depleted by 2 m during these 24 h (without any rainfall event). Previous pumping tests of this well resulted in a transmissivity value of 1×10^{-3} m^2/s. Considering that the well is located at the edge of the catchment and without any inside–outside lateral flow, what is the storativity value for the aquifer section which becomes completely drained between the well and spring? (Fig. 3.23)

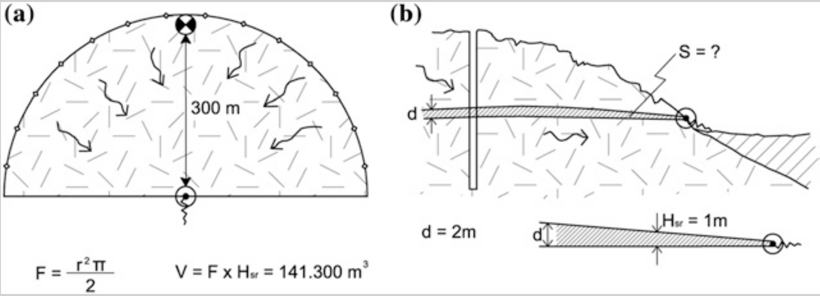


Fig. 3.23 Sketch map and cross section within the spring's catchment

By applying formula (3.12), we calculate storativity to be:

$$S = 2.25 \times T \times t/r^2;\ S = 2.25 \times 0.001 \times 86,400/300^2;\ S = 0.00216(0.22\,\%)$$

By applying formula (3.13), we obtain as follows:

$$S = V_t/V;\ S = 0.0035 \times 86,400/141,300 = 0.00214(0.22\,\%)$$

Obtained storativity rates specify an aquifer which is not highly karstified and with modest storage. The fissure porosity in the studied part probably prevails over the cavernous one. Secondly, although the results of the two equations are relatively similar, this is not always the case and these approximation formulae have to be compared with results of field surveys and several methods suitable for an assessment of aquifer storage (Stevanović et al. 2010).

Box 3.8

Castany (1984) presented results of calculated storativity and transmissivity obtained from infield surveys and tests undertaken in a variety of karstic aquifers in central Europe and in the Mediterranean region (Table 3.1).

Table 3.1 Storativity and transmissivity for some carbonate karstic aquifers (reproduced from Castany 1984)

Carbonate rock category	Age	Location	Storativity S (%)	Transmissivity T (m^2/s)
Fissured limestones	Upper Jurassic	Moutier (Switzerland)	1–1.5	
	Turonian–Cenomanian	Israel	1	$0.1 \times 10^{-2} - 1.3 \times 10^{-1}$
	Upper Cretaceous	Tunisia	0.5–1	
	Miocene	Murcia (Spain)	0.7–1	
	Jurassic	Lebanon	0.1–2.4	$0.1. \times 10^{-2} - 6 \times 10^{-2}$
Karstic-fissured limestones	Lias	Tunisia	4–5	
	Upper Jurassic	Tunisia	5–7	
	Urgonian (lower cretaceous)	Salon (France)	1–5	10^{-3}
	Jurassic	Parnassos (Greece)	5	$1–2 \times 10^{-3}$
	Jurassic	Vaucluse (France)	1–5	
	Jurassic	Grand Causses (France)		10^{-2}
Fissured dolomite	Jurassic	Grand Causses (France)		10^{-3}
	Lias	Morocco		10^{-2}–10^{-4}
	Jurassic	Parnassos (Greece)		3×10^{-5}
Fractured marble		Almeria (Spain)	10–12	
Fissured dolomite		Murcia (Spain)	7	
Marly limestone	Jurassic	Grand Causses (France)		10^{-3}

Referring to storativity, in many karst studies there is a statement of the two dominant types of karst systems (Atkinson 1977; Bonacci 1993; Padilla et al. 1994; Panagopoulos and Lambrakis 2006): (1) Poorly developed karst systems with large storage capacity (diffuse) and (2) well-developed karst aquifer with larger conduits but without significant storage of water. In contrasting these two systems, large phreatic reservoirs in highly karstified limestones, such as those of Yucatan, are neither evaluated nor classified often (Stevanović et al. 2010).

Therefore, there are many karstic aquifers with large effective porosity, permeability, and storativity where due to a large and generally full reservoir, the impulse of newly infiltrated water is slowly transferred to discharge point. This is typical for deep siphonal karstic aquifer drained by ascending springs. Although it is not the same as, it is somehow similar to the reaction of a large surface reservoir on an increased river flow upstream of a water tail (of course, not dramatically increased), impulse will slowly transmit until it reaches the last point—the dam. Advantages of such type of drainage and large karstic reservoirs for engineering regulation are discussed in Sect. 15.5.

3.3 Flow Types and Pattern

Ten years before Henry Darcy did his survey in Dijon and formulated his famous law, Poiseuille's (1846) studied water flow through small tubes and found that *specific discharge* as flow per unit of a cross-sectional area is directly proportional to the hydraulic head loss between one end of the tube and other (Ford and Williams 2007). Poiseuille imposed gravitational acceleration, fluid density, dynamic viscosity, and hydraulic gradient as additional impact factors influencing specific discharge rate.

Darcy's law assumes *laminar flow*. This approximates the constant diameter of the tube or pore system which the water is flowing through. Hypothetically, the water particles move in parallel threads in the flow direction (non-disturbed parallel streamlines). *The turbulent flow* appears when the velocity and diameter of tubes increase and particles start to fluctuate with transverse mixing (streamlines are disturbed). The Reynold's number (R_e) identifies the critical velocity. It depends on the diameter of the tube (d), fluid velocity (v), and the two factors imposed by Poiseuille: fluid density (ρ) and dynamic viscosity (μ):

$$R_e = \rho \times v \times d/\mu \tag{3.14}$$

Some other standard hydraulic formulae are also applicable for specific cases of groundwater flow in karst and many mathematical deterministic models include equations of Bernouli, Chézy, and Maning. Details can be found in reference literature.

Krešić (2013) noticed that the traditional hydraulic of tubes (pipes, channels) is based on the principle of flow continuity (Fig. 3.24), which assumes that there is no inflow or outflow through the tube's walls. The elementary flow tube (Q) is directly proportional to the elementary cross-sectional area (A) and the average flow velocity (V_{av})

$$Q = A \times V_{\mathrm{av}} \quad (\mathrm{m}^3/\mathrm{s}) \tag{3.15}$$

Krešić (2013) also stated that ideal tubes are rare in karst aquifer and that an extensive exchange of water between main conduits ("tubes") and the rock matrix usually takes place. Influx or outflux in the same conduit depends on pressure, and if the conduit is fully saturated, it discharges into the matrix. During the recession periods, the process is reversed, and water from the matrix discharges back to the conduit with a free water surface (Fig. 3.25).

Krešić (2007) highlighted the four factors which complicate calculation of flow through natural karst conduits:

- Inconsistency of pressure in the same conduit (presence of segments under pressure and with free surface);
- Irregularity of conduit's walls which requires estimation of the coefficient of roughness (and has to be inserted into the flow equation);

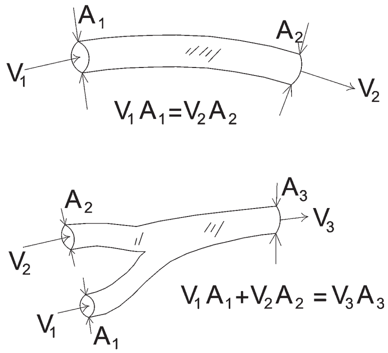


Fig. 3.24 Flow tubes and principle of flow continuity (after Krešić 2013, reprinted with permission of McGraw-Hill Education)

Fig. 3.25 Temporary dried cervical channel ("tube") in Devonian carbonate matrix (Zengpiyan, Guilin, China) (*left*) and active turbulent flow in Stopića cave (Zlatibor, Serbia) (*right*)

- Variability of cross sections along the conduit (even within short distances); and
- Changeability of flow types (laminar and turbulent) in the same conduit, caused by factors such as cross-sectional area, wall roughness, and flow velocity.

Finally, determination of the position of the main conduit (if there is just one preferential) is sometimes considered to be the main target of a survey (whatever the task is: water extraction or prevention of leakage) and a positive result heightens the possibility of having a successful project (Box 3.9). Bakalowicz (2005) emphasizes geophysics and field mapping of fracture distribution as generally efficient enough to locate drillings in productive zones. Goldscheider and Drew (2007) edited a book on different methods applied in karst hydrogeology which can, along with book of Petar Milanović's (1981), be used as a reference to the topic.

Box 3.9

The borehole HG-31 was drilled under the investigation program for the regional water system "Bogovina" in eastern Serbian karst (Stevanović 2010a). The location was chosen, as it was for all other exploratory boreholes, based on results of extensive infield survey. The logging and airlifting of HG-31 fully confirmed the choice: During the drilling at an interval of 77.6–98.2 m, a very small percentage of cores were removed and this part was drilled very fast. Analyzed core fragments were dark gray and round, which was supposed to be a possible indicator of active water circulation near an edge of an open cavernous zone. A continual yield of 2 l/s with a stable level in the small diameter piezometer HG-31 during the airlifting was also an indicator of promising well productivity. Considering these positive results of borehole HG-31, a good yield from well planned at that location was expected. The design of extraction well IE-5 was similar to the others drilled in the well field (Q_{av} = 40 l/s) and included an open-hole interval with a diameter of 295 mm in the deepest part of the well, at a depth of 59.5–110 m (Fig. 3.26).

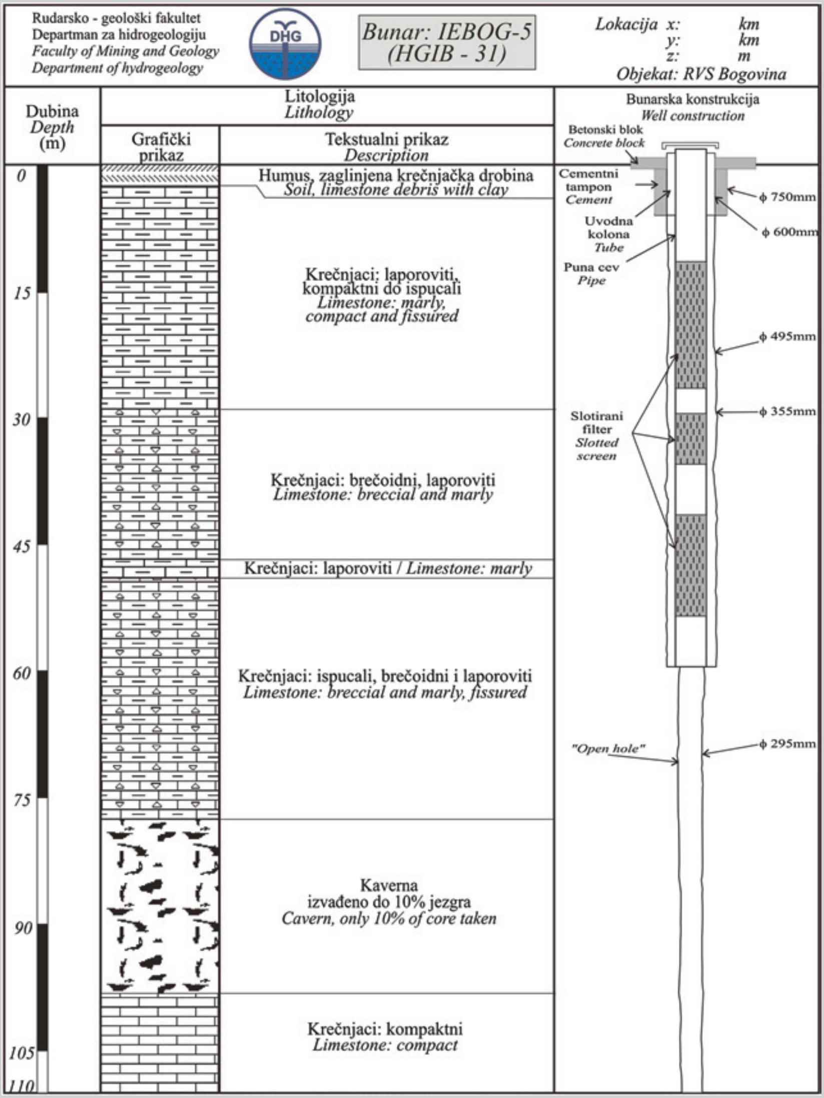


Fig. 3.26 Design and log of well IE-5 (Bogovina, Serbia; from Stevanović 2010a)

Unfortunately, this well is not as productive as expected. Its discharge is 15 l/s per drawdown of 21 m. The smaller capacity is most probably the result of having moved the well location by just 1 m from the HG-31 borehole. Although such situations have happened elsewhere in karst, previous positive experience with other wells and the wish to keep piezometer HG-31 as an important nearby observation point prevailed and brought about an undesired result. However, a good lesson was learned. Later, a remediation project including controlled explosions within the open-hole interval was prepared in order to open the route between the two holes, but after long discussion, this solution did not come into effect (mostly because of sufficient productivity of other wells).

White presented a classification (1969, and updated later) of flow types in carbonate aquifer system in regions of low-to-moderate relief: (1) *Diffuse flow* characterized shaley limestones and crystalline dolomites with high primary porosity or uniformly distributed fractures; (2) *Free flow* is linked to thick, massive soluble rocks, both perched and deep unconfined aquifers with conduits developed along bedding, joints, fractures, or fold axes; (3) *Confined flow* is diffuse or free flow and takes place in artesian and sandwiched aquifers overlain by beds of low permeability.

By analyzing the epikarst flow to the underlying vadose zone, the seepage flow in the upper parts has been differentiated from vertical drainage through a large fissure in the zone of lower aquifers (Fig. 3.6). Vertical drainage is the result of the connection of numerous small seepages which may reflect *diffuse flow* into one or a few vertically percolated flows. The latter can be termed *concentrated flow*. These conditions again certify the dual or multi-porosity of a karst aquifer. Finally, the part of lowest aquifer in the zone of saturation of unconfined aquifers is characterized by movement of water under described hydraulic laws but in principle with a water table (potentiometric pressure) slightly inclined toward discharge points or base of erosion. Such conditions were described by Cvijić in his theory of water circulation (1918). His coherent synthesis introduced three main superpositioned zones in specific dynamic coexistence. Vertical gravity circulation is dominant in the uppermost part, while the deeper saturated zones are characterized by (sub)horizontal and/or ascending circulation. Cvijić concluded that permanent lowering of the saturated part of the carbonate rocks is a logical consequence of the dynamic evolution of the karst (Fig. 3.27).

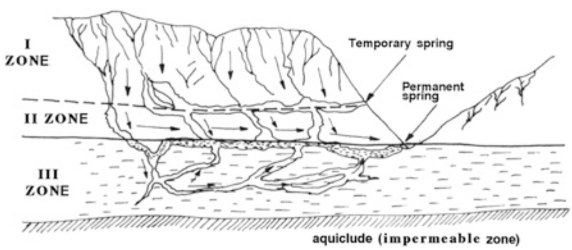


Fig. 3.27 Cvijić's concept of three "hydrographic" zones in karst: *I zone*—"dry" with percolation only; *II zone*—"transition" zone of water table fluctuation causing both vertical percolation but also gravity discharge; *III zone*—"stagnant" zone, fully saturated with siphons and conduits with ascending flow

Mangin (1975) suggested the existence of two types of flow through the vadose zone. Figure 3.28 shows that percolation of infiltrated rain or water of sinking streams might be *slow* or *fast* depending on aquifer properties.

This is not very different from the previous separation of *diffuse* and *concentrated* flow.

Another classification considers flow hierarchy as defined by Tóth (1999, 2009). Figure 3.29 shows the general concept of Tóth's theory and superpositions of gravity-driven flows: *regional flow, intermediate flow,* and *local flow*. They are characterized by the functionally different flow regimes superhydrostatic, hydrostatic, and subhydrostatic, respectively. Application of this theory in deep structures and confined aquifers is discussed in Sect. 17.5.

It is clear from the below scheme that the system of karstic conduits could also be superimposed even in the same hydraulic zone. Therefore, a network of karstic paths which enables water transfer may be vertically but also horizontally arranged. In many field experiments, the main or "chief" conduit has been identified, while others have a subordinate role (Milanović 1981). The main conduit is often termed *principal* (*dominant*), while the others are *peripheral*. Figure 3.30 shows the result of a tracing experiment and curve tracer concentration versus time. While the

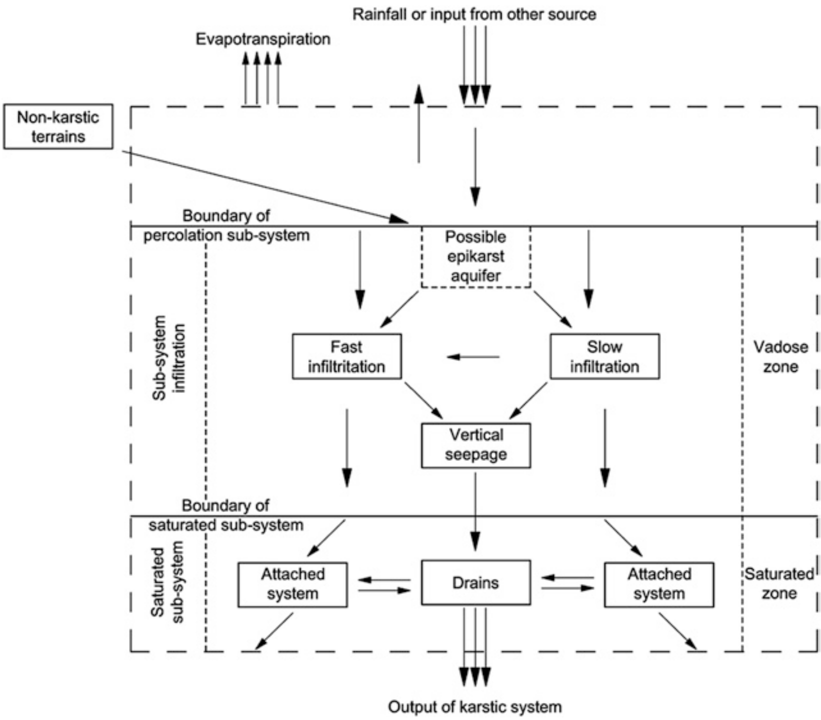


Fig. 3.28 Scheme of a karst aquifer function (from Mangin 1975)

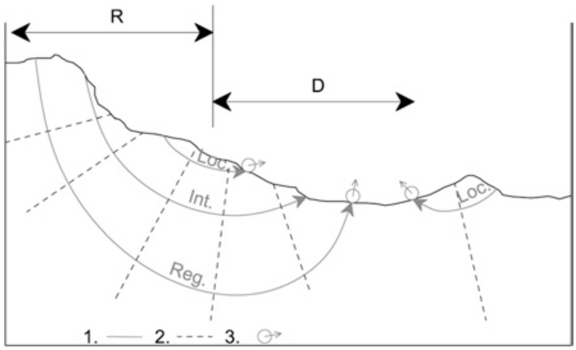


Fig. 3.29 Conceptual interpretation of Tóth's solution for gravity-driven flow in unconfined drainage basin. *Legend 1* streamline (flow line), *2* equipotential line, *3* spring, Reg/Int/Loc—regional, intermediate, and local flow, *R* recharge zone, *D* drainage zone

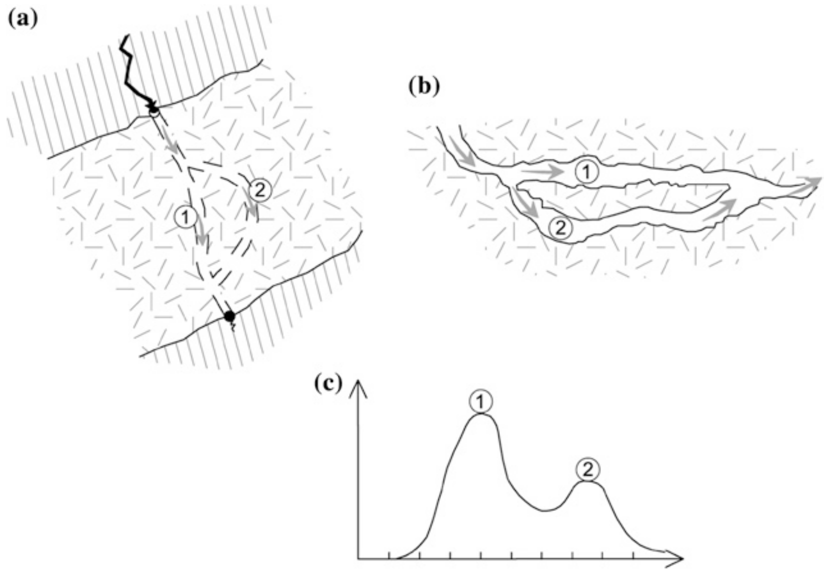


Fig. 3.30 Principal (*1*) and peripheral (*2*) channels and flows connecting ponor and spring. **a** Map—lateral difference. **b** Cross section—vertical difference. **c** Tracer concentration curve

first and higher peak indicates the time when the tracer was first recorded passing through the main conduit, the second peak is a result of peripheral flow.

The replacement of roles of superimposed conduits happens often in karst. When during the recession periods pressure in an aquifer decreases and water from pressurized cavities ("tubes") starts to flow freely instead, and the upper cavity has dried out, the peripheral bottom conduit assumes the main role in water transportation to the discharge point. Cases of groundwater piracy and altering of the flow to the “new” catchment, or changing directions of flow throughout the year, are similar (Box 3.10). The final stage of evolution of karst aquifer considers adaptation of water flow to the base of erosion. All transition occurrences may be considered as incidental indicators of the karstification process, which is ongoing. This is the case with the dual function system of estavelles (spring ponor), or temporary aquifer piracy, or conversion of a flow’s direction during a limited period of time.

Box 3.10

Figure 3.31 shows a characteristic subterranean hydraulic mechanism in the Kučaj karst massif in eastern Serbian karst. At higher water periods, drainage takes place at two discharge points (Mrljiš and Fundonj springs), both at the edge of the karst massif but dislocated in two directions. Such a mechanism has been confirmed by the tracing of the Bogovina cave ponor (tracer simultaneously appeared in both springs). During the recession periods, the decline in the potentiometric pressure allows flow only to the regional base level of erosion in the area—to the Mrljiš spring and the Crni Timok River valley (Stevanović 2010a).

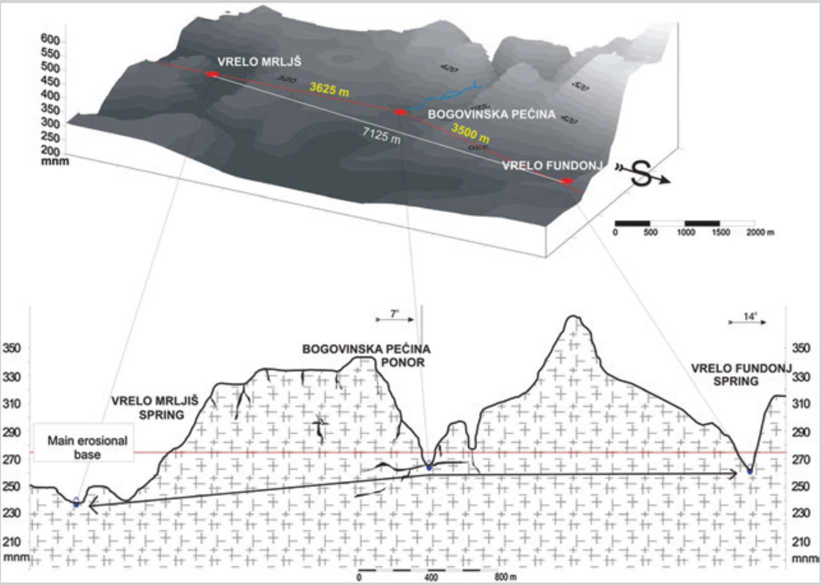


Fig. 3.31 Two-directional flow from ponor zone of Bogovina cave converts into one-directional flow during the recession periods (from Stevanović 2010a)

3.4 Aquifer Recharge

Two main types of recharge exist: *natural* and *artificial*. The main focus here will be on the natural recharge processes.

Lerner et al. (1990) describes natural recharge as the flow of the infiltrated water which reaches the groundwater table and results in an increase of water volume stored in aquifer.

The natural recharge sources are as follows:

- Precipitation,
- Surface waters: sinking streams, lakes, and sea water,
- Underground flows from adjacent catchment.

1. *Precipitation* includes rain water, snow melting, and condensate water. In most of the continents between 60°N and 60°S, except on high mountains rainfall is the dominant recharge component and its contribution is dispersed throughout a whole year, while *snow melting* is relatively short and regularly results in high peaks in the springs' hydrograph. *Condensation* also requires differences in air and soil temperature and humidity and usually contributes the least to replenishment of an aquifer's water reserves.

The infiltration of rainfall and snowmelt, including condensation to a certain level, all produce the earlier discussed *diffuse flow*. In fact, the character of flow is more dictated by soil cover or tectonic fabric and fissures' aperture, but, however, small the seepage of rainy water, it cannot be neglected as a factor. The exception is very high rainy storm episodes or intensive and fast snow melting when diffuse flow may convert into *concentrated flow* (Box 3.11).

Box 3.11

An example of the temporary functioning of a small sinking brook resulting from snow melting was described by Raeisi (2010). Thirty kilograms of sodium fluorescein were injected into this sinking brook ending in a nearby ponor (swallow hole) in the northern flank of Barm-Firooz anticline (80 km NW of Shiraz, Iran) (Fig. 3.32). The sampling sites were more than 30 springs of the two karstic formations Sarvak Fm. (Cretaceous) and Asmari-Jahrum Fm. (Paleogene), but a tracer only appeared at 18 km from the Sheshpeer spring.

Four experiments were performed on the catchment area of Sheshpeer Spring from December 1991 to April 1992 to determine the effect of external parameters such as flow rate, specific conductivity, temperature, and dissolved ions of recharged water on the physicochemical characteristics of the Sheshpeer spring. In the second experiment, snowmelt flow was measured every 2 h for 6 days. The ratio of average daily maximum flow rate to the average daily minimum flow rate varied from 5 to 10. However, no effect of such a daily oscillatory recharge had been observed on the Sheshpeer spring hydrograph. A distance of 15 km between input and output points seems to be enough to suppress the effect of daily oscillations of the input flow rate.

Fig. 3.32 Karrenfeld in the core of Barm-Firooz anticline (Iran). Small and short brooks form only during the snow melt

2. *Concentrated recharge* also called *point recharge* is typical for the waters which infiltrate from *sinking streams* and concentric ponors. The ponor capacity is an equivalent of spring discharge but with inverse function. It is a volume of water in a unit of time which may be absorbed by the rocks. Some ponors are really huge and may swallow and transfer yields equivalent to thousands of m^3/s of water, but in principle, such yields are rare nowadays. Therefore, big cave ponors (Fig. 3.33) played a more evident role in the intensity of the karstification process in the past.

In contrast, many sinking streams are characterized by very small apertures in their riverbeds and the only means to identify the presence of sink streams (swallets) is to conduct simultaneous hydrometry at consecutive sections along the riverbed and to find evidence of the water losses (see Chap. 6).

As a result of aquifer's high permeability, recharge of karstic aquifer by lake water or undesired sea water is frequent in coastal aquifers (see Sect. 16.4).

3. *Underground flows from adjacent aquifers* and catchment are not often discussed in karst literature. Most authors consider this as simply a flow continuum, but for aquifer budget and assessment of water reserves, this is a very important issue. The lateral inflow or seepage from overlying aquifers (Fig. 3.34) may be continual, and then, revision of the basin geometry is probably required. But seepage may also take place for a limited period depending on potentiometric pressure and water table fluctuation in both receiving and emitting aquifers.

Fig. 3.33 Active cave ponors in the Carpathian and Dinaric karst (Sohodol, Runcu gorge, Romania, *left*) and (Rakov Škocjan, Slovenia, *right*)

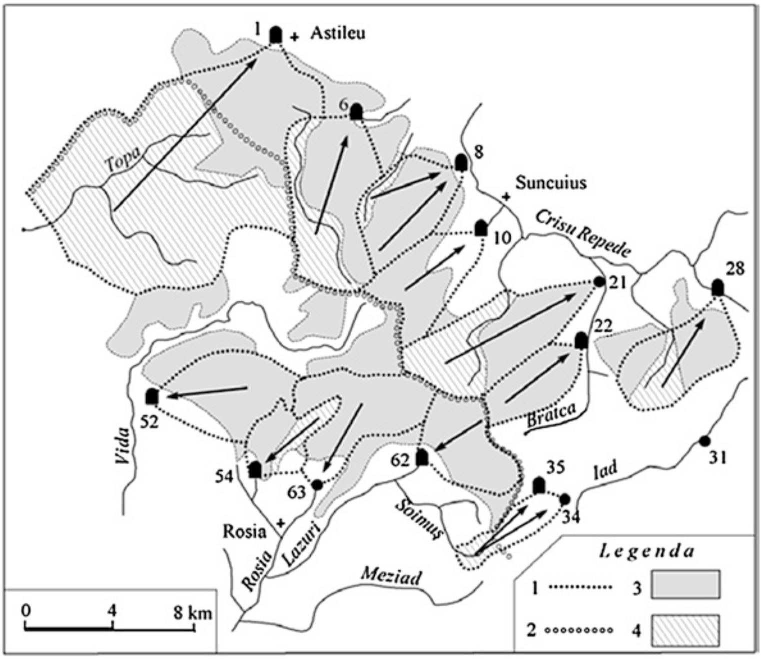


Fig. 3.34 The sketch map of Padurea Craiului Mt. (Romania): Karstic aquifers with a large extension of non-karstic terrains contributing to allogenic recharge. *Legend 1* groundwater divide, *2* watershed divide, *3* karstic terrains, *4* non-karstic rocks. Reproduced by permission from Oraşeanu and Iurkiewicz (2010)

When infiltration results from water formed inside a delineated karstic catchment, it is considered as *direct recharge* or *autogenic recharge*. If water for infiltration arrives from non-karstic terrains and contributing catchments, *indirect recharge* or *allogenic recharge* is spoken about. The system consists exclusively from karstic rocks with autogenic recharge is also called *unary karst system*, while *binary karst*

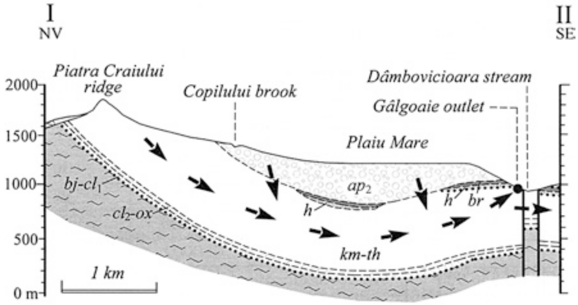


Fig. 3.35 Model of Dambovicioara passage (Padurea Craiului Mt. Romania). Vertical seepage is from overlying recent deposits (*gray*) into the main karstic Mesozoic carbonate karstic aquifer (*white*). Reproduced by permission from Oraşeanu and Iurkiewicz (2010)

system includes non-karstic rocks and their contributing catchment (Marsaud 1997; Bakalowicz 2005).

Many still active ponors are located directly at or near the contact of karst and impervious or low permeable rocks which convey perennial streams toward those aquifer recharge points (Fig. 3.35). There are also many blind dry valleys formed at such contacts providing evidence of karstification and enabling reconstruction of the paleo-hydrographic situation.

Effective infiltration as the amount of water which will efficiently reach the water table and storage (Box 3.12) depends on many factors, and in the case of water originating from the atmosphere involves climatic and water balance elements such as rainfall intensity, evapotranspiration, wind and other elements discussed in Chap. 6. However, along with these climatic factors there are some other factors which may be grouped as follows:

3.4.1 Non-geological Factors of Natural Recharge

Topography: Slope is one of the main factors, depressions such as uvala or dolines regularly favor water stabilization and infiltration, while higher slopes stimulate a runoff component. Therefore, terrains with smaller slopes are more suitable for groundwater recharge. Digital elevation models, topography maps, and remote sensing support slope analysis.

Soil cover: If this is missing, the infiltration will be more effective. If it is present and thick, then it regularly reduces efficiency of infiltration (Fig. 3.36). Pedology maps and digital Corine land cover maps in combination with remote sensing may facilitate this assessment.

Soil moisture: When soil cover is present then moisture balance becomes an important factor for infiltration rate. If soil is fully saturated from previous rains, melted snow or flooding then runoff and evapotranspiration will again be dominant in the water balance, and less water will infiltrate. But in arid karst or during

Fig. 3.36 The vertical section of composite carbonate—gypsum layers prosperous for active recharge from soil cover and moisture point of views: soil is very thin and layers are dry due to arid environment (Taalex Fm. in Laasqoray Xudun—Sool Plateau, Somalia)

the long and dry seasons when soil moisture deficit grows, macro pores and cracks are activated causing the infiltration capacity also to rise (Fig. 3.36).

Vegetation: Plant distribution and density also play an important role in the water cycle. The grass, cultivated crops, and trees are all extensive water consumers, and if those are extended and dense, the transpiration will increase causing a reduction in water available for infiltration. Moreover, some experiences show that during the vegetation seasons in old and dense forests, rains of little intensity with a sum of less than 3 mm/day will not reach the land and will not have a chance to infiltrate. *Interception* represents the difference between the total precipitation which falls and the precipitation that actually reaches the soil; thus, water is intercepted by the leaves and plants.

3.4.2 The Geological Factors Influencing Recharge

Lithology: Not all dissoluble rocks have the same infiltration capacity; those more soluble and with greater primary porosity such as the evaporitic group will infiltrate more water. Overlying layers such as thick overburdening impervious rocks and lithological impurities also have a great impact causing reduction of inflow.

Tectonic fabric: It is probably the central geological factor. If surficial fissuration is very small or missing and the bedding plane is horizontal, then infiltration will not take place or will be minimal. In fact only small intensity surficial karstification will then take place. Such a situation is presented on Figs. 2.12 and 2.13. In contrast, both epikarst and bare karst with developed vertical cavities are suitable environments for intense infiltration. Tectonically fabricated vertical or subvertical strata also support water infiltration between bedding planes (Fig. 3.37).

Aquifer saturation: If aquifer is completely full of water, and the water table reaches land surface, no new water can be added to the ground and then runoff prevails.

To conclude, Fig. 3.38 illustrates the explained components of natural recharge of a typical unconfined karst aquifer. The rainfalls produce mostly diffuse recharge

Fig. 3.37 Structures suitable for receiving and conveying waters: bedding plane aperture (Devonian karst of south China, *left*), widely open fractures (Eramosa dolomitic karst, Ontario, Canada, *center*), and highly inclined carbonate rocks (Sierra de la Nieves, south Spain, *right*)

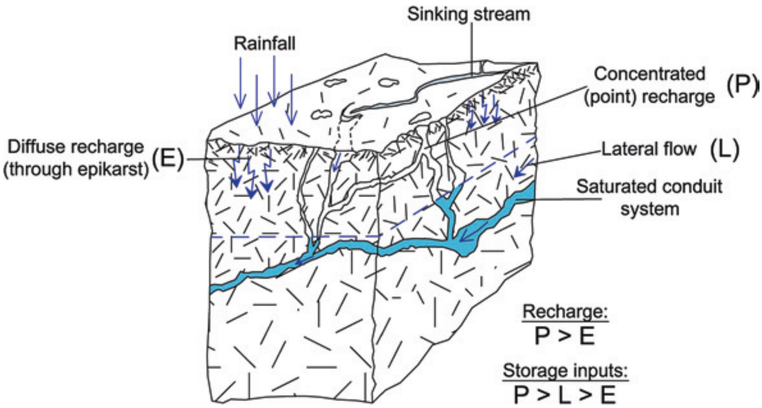


Fig. 3.38 Schematic model of natural recharge components in an unconfined karstic aquifer

and slow percolation through epikarst (*E*). Sinking streams lose all their water in a big concentric ponor which extends to a great depth in the form of a pothole enabling concentric water flow to the water table (*P*). Finally, lateral flow (*L*) conveys water from a remote part or from another aquifer or adjacent basin. When vertical cavities are sufficiently large to convey sunken water, the point recharge is higher than diffuse through epikarst: $P > E$, and similarly $P > L > E$.

The term *artificial recharge* is often improperly identified with *engineering regulation of aquifer*. Although in intergranular media this is justifiable because the majority of interventions consider precisely the artificial transport and induced inflow (whether with or without pressure) into the aquifer, this is not often a solution applied in karst. There is simply a general awareness that water in karst may "escape" through many unknown corridors and that effects of such interventions would be minimal. But similar interventions with regulating the riverbeds or redirecting surface streams to recharge ponors in other basins are known in engineering practice (see Chap. 13). However, the artificial recharge of aquifer in platform karst is more promising and recharge wells are being used in Florida to convey waters from shallow to the underlying Floridian karstic aquifer.

Finally, artificial recharge may take place unintentionally. Water use in irrigation can lead to substantial increases in recharge. This may be the indirect result of soil watering or leakage from irrigation conveying channels. Younger (2007) pointed out that not only seepage from water pipelines and septic tanks, but also intensive watering of parks and gardens increases recharge in Ryadh, Saudi Arabia. Such recharge is commonly termed as *irrigation return flow*.

Areas urbanized on karst are ideal places to have significant recharge of aquifers. Knowing that losses from water-conveying pipes presented by water utilities are as great as ca. 10 % to as much as 60–70 %, which is occurring in some undeveloped cities, it is clear that a large portion of lost water may re-infiltrate in the aquifer. Sharp and Garcia-Fresca (2004) estimated the drinking water leakage in Austin, Texas, to be around 7.7 % or 21 mm/year as water available to be re-infiltrated into the Edwards aquifer. The amount is deemed significant considering that natural recharge of local aquifer is in the range of 30–100 mm/year. An example of incidental recharge of karstic aquifer is also presented in Sect. 3.6, Box 3.18. Nevertheless, seepage from waste water reservoirs and channels, storm sewers, or septic tanks is also a common recharge source in urban areas.

From above, it is clear that effective infiltration is variable element. If we thus express it as % of precipitation as common in hydrology practice, we should consider the time component, i.e., period in which effective infiltration is taking place. For a rough and general assessment of groundwater recharge, an average annual value may be acceptable as presented in Table 3.2 (Box 3.12).

Box 3.12

Table 3.2 Estimated effective infiltration rates of some karstic aquifers

Type of aquifer rock category	Location	Effective infiltration in % of annual precipitation (%)	Author	Source (reference)
Cretaceous chalk	London basin, England	20–35	Anon, 1972	Lerner et al. (1990)
Paleozoic dolomites	Ghaap Plateau, South Africa	2–25	Smith, 1978	Lerner et al. (1990)
Taalex gypsum, Eocene	Ceerigaboo, Somalia	30	Stevanović et al. 2012a	–
Pliocene–Pleistocene limestones	Morroco	14–19.5	Bolelli, 1951	LaMoreaux et al. (1984)
Aptien-Albien limestones	Dj. Sidi Rheriss, Oum el Bouaghi, Algeria	38	Stevanović, 1985	–
Eocene limestones	Dyr el Kef, Tunisia	33.2–90.0	Schoeller, 1948	LaMoreaux et al. (1984)

(continued)

Table 3.2 (continued)

Type of aquifer rock category	Location	Effective infiltration in % of annual precipitation (%)	Author	Source (reference)
Senonian and Jurassic limestones	Dj. Chounata, Tunisia	33–53	Tixernot et al. 1951	LaMoreaux et al. (1984)
Middle Cretaceous limestones	Na'am spring basin, Israel	53	Mero, 1958	LaMoreaux et al. (1984)
Pila Spi Fm., Tertiary limestones	Dohuk, northern Iraq	35	Stevanović, 2003	–
Bekhme Fm., Cretaceous limestones	Harir, northern Iraq	40	Stevanović and Iurkiewicz, 2004	–
Mesozoic limestones	Lilaia springs, Greece	51.6	Aronis et al. 1961	LaMoreaux et al. (1984)
Mesozoic limestones	Monte Simbriuni, Italy	69	Boni and Bono, 1984	Burger and Dubretret (1984)
Mesozoic limestones	Monte Lepini, Italy	78	Boni and Bono, 1984	Burger and Dubretret (1984)
Jurassic and Cretaceous limestones	Kučaj-Beljanica Mts. east Serbia	31	Stevanović, 1991	–
Triassic limestones	Tara Mt. west Serbia	47	Stevanović, 1995	–
Mesozoic limestones	Malé Karpaty, Slovakia	24.4–52.4	Kullman, 1977	Burger and Dubretret (1984)

Numerous methods are applied to estimate recharge of a karstic aquifer; some of them are discussed in Chaps. 4–6 and other contributions to this book. Among them are groundwater budgeting, stochastic input–output modeling, isotopic methods, Cl^- ion balance, and GIS application methods.

3.5 Aquifer Discharge

Drainage of karst aquifer is a crucial process for most planned engineering works, whether we intend to tap and utilize karstic waters or defend ourselves from them. Therefore, the majority of research activities in karst hydrogeology usually concentrate on the discharge points as the places where direct access to issuing groundwater is possible. In many cases, concentration of works near discharge points is also dictated by the topography and accessibility, and while many spring sites are in reachable foothills or in flat areas, their catchments could be in high mountains not easily reached, glaciers or dense tropical forests.

Drainage of karst aquifer or aquifer's discharge takes place as (Stevanović 2010b) follows:

Natural, carried out through springs, submerged springs, estavelles, subsurface outflow and

Artificial, effectuated through groundwater extraction by wells, galleries, canals, or similar water intake structures.

The topic of artificial drainage and tapping karstic waters is discussed in Chap. 11. From the point of view of water budget, natural drainage of aquifer also includes a few other components such as runoff from karstic surface or evapotranspiration, both discussed in Chap. 6. The subsurface outflow as an important way of drainage, even essential for many engineering works in karst, is also discussed in Chap. 6 and Sect. 15.5. What follows is thus focused on natural discharge and interrelated factors.

The intensity and velocity of groundwater flow toward discharge points depend mostly on the hydraulic gradient. The greater the inclination of the water table and the larger the hydraulic head, the greater the energy that will push stored water to flow out from the aquifer (Fiorillo 2011). The limitations on effectual drainage can be the opening of discharge point(s), i.e., *orifice,* and its dimensions which may not allow quick discharging or even total emptying of an aquifer. When the water table and hydraulic head decline over time, the discharge rates also decline usually in dry seasons, but it sometimes takes a long time before the aquifer completely dries out (Fig. 3.39). The *coefficient of discharge,* also called the *recession coefficient* (α), which expresses the diversity of an aquifer's characteristics and the storage capacities which influence discharge intensity, is explained in Chap. 7.

Groundwater flow aims to reach the lowest discharge point of an aquifer which commonly corresponds to the local or regional erosional base. *Erosional base* or *base level of erosion* is the lowest level of the terrain where erosion by water is still possible. The erosional base is changeable in geological time, and the evolution of the karstic process follows and adapts to that descending level. Many dry karstic caves had previously functioned as springs and superpositioned big cave openings carved in karst at the cliffs in many canyons provide evidence of this process (Fig. 3.40).

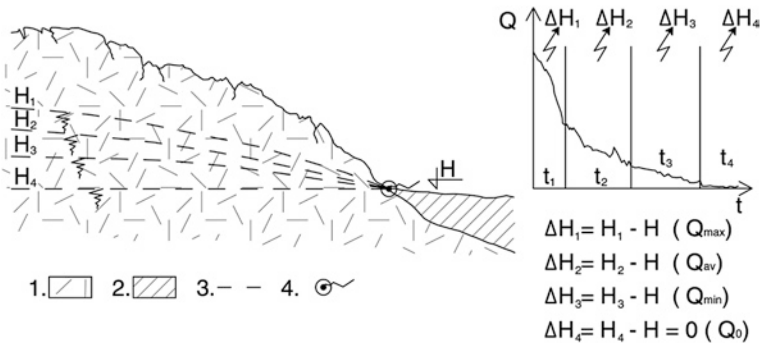


Fig. 3.39 Discharge intensity depends on hydraulic head. The maximum discharge is during the period t_1 when the hydraulic head has maximal value (ΔH_1), while during the period t_4, the spring is running dry ($\Delta H_4 = 0$, $Q = 0$). *Legend 1* karstic aquifer, *2* impervious rocks, *3* water table or hydraulic head, *4* spring, *Q* discharge, *t* time

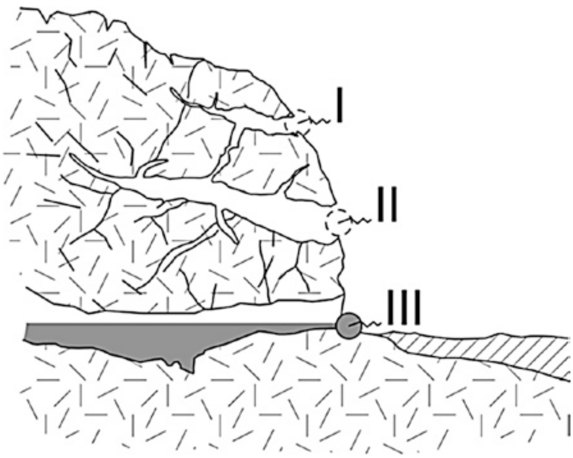


Fig. 3.40 Evolution of karst aquifer and adaptation of discharge points to the erosional base. *I*, *II* previous drainage points at the cliff; *III* actual deep ascending discharge point

The erosional base is a principal controlling factor of the development of karstic landforms (Ford and Williams 2007) and of groundwater deepening. However, the *erosional base* cannot be equaled to the *karstification base,* and in mature karst, the process of karstification and rock dissolution may take place deep below the actual erosional base.

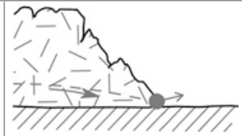
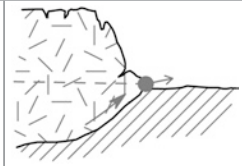
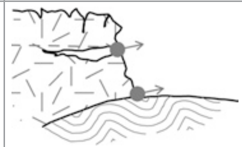
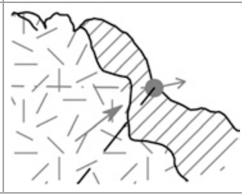
The *regional base level* is equivalent to major rivers, lakes, or seas and zones to which the regional groundwater flow tries to accommodate. There are also *local erosional bases*, which may be represented by a lithological barrier or non-permeable layer within an aquifer, fault, anticline plunge, depression, or other geological and topographical elements, which influence and predispose groundwater drainage. One such local erosional base is shown in Fig. 3.2 (see C: case of perched aquifer).

A spring is a natural opening (*orifice*) of the land surface from which groundwater visibly flows out from an aquifer. Thanks to springs and their usually sufficient openings, karstic aquifers mostly discharge concentrated flows; diffuse flows are definitely present in karst to a lesser extent than in any other aquifer system.

Many authors afforded classifications of the springs as outlets of aquifers. Along with the description and explanation of the function of some major springs in the USA, Meinzer (1923b) distinguished categories in accordance with springs' discharge rates. Bögli (1980) provided a very comprehensive and well-explained classification which includes several groups of springs. Herak et al. (1981) stated that characteristics of springs are defined by the properties of aquifer, the position of connected aquicludes and aquitards, and by the climate and vegetative cover. Ford and Williams (2007) also explained the main spring types and distinguished the three major groups (free draining, dammed, and confined). In addition, they provided a list of the world's largest springs. Krešić (2010) wrote a chapter on types and classification of springs in the textbook "Groundwater hydrology of springs," while in the same book, Stevanović (2010b) focused on utilization and regulation of springs based on their characteristics.

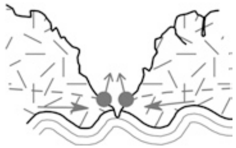
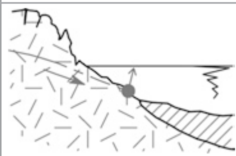
There are several classifications of springs made in accordance with established criterion. Table 3.3 is an attempt to present some of those criteria, ten in total, and accordingly to provide common classifications and typology of springs.

Table 3.3 Classification of springs (criteria and typology)

Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
1. Flow type and pressure in aquifer	1.1 Descending	Gravity	Free draining aquifer. Aquifer with water table (phreatic) inclined toward discharge point: gravity downward flow	
	1.2 Ascending	Artesian, confined, raising	Confined aquifer. Water driven by pressure: upward flow	
	1.3 Overflowing		Usually free draining aquifer. Often upper aquifer's outlet functioning periodically. Slightly inclined or subhorizontal water table: flow pours over barrier	
2. Barrier position and type	2.1 Contact	Dammed, bedding	Lithologically predisposed. Water issuing at the lithological contact with non-karstic rocks of lower permeability (total barrier)	
	2.2 Fracture	Fault	Tectonically guided flow. Water issuing from an open crashed zone, fracture, fault or other lineament	

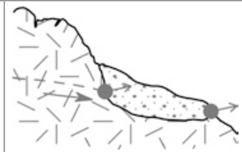
(continued)

Table 3.3 (continued)

Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
	2.3 Hanging		Lithologically or tectonically predisposed local barrier. Water issuing above main (regional) erosional base. Often drainage of perched aquifers	
	2.4 Anticlinal		Tectonically predisposed. Water issuing from anticline plunge (margin of the structure)	
	2.5 Depressional		Predisposed by topography and more often by river incision. Depression in relief intersects water table enabling free flow from aquifer. The same is true with cutoff of confined aquifers	
	2.6 Impounded	Subaqueous, submarine, sublacustrine, submerged, vrulja	Discharge point is impounded by fresh, lake, or river water, or is under the sea level (totally or intertidal). The karstification is thus below the erosional base and barrier may be absent	 See also Box 3.13

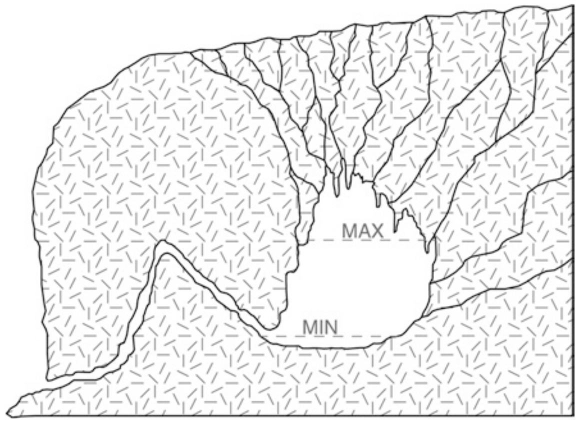
(continued)

Table 3.3 (continued)

Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
3. Spring–aquifer relation	3.1 Primary		Water issuing directly from original karstic aquifer	As in case 1.1–1.3
	3.2 Secondary	Masked	Water issuing from connected lateral aquifer or from rock debris away from original discharge point	
4. Discharge point feature	4.1 Cave		Water issuing from a hydrologically active cave	See Fig. 3.41i
	4.2 Siphonal		Water issuing from siphon or from siphonal reservoir, usually as an ascending flow	See Fig. 3.41j
	4.3 Lake	Vauclusian	Water issuing from the lake usually extends to depth as funnel-like vertical channel formed by erosion of a strong ascending flow. The name of the famous French spring Fontaine de Vaucluse (near Avignon) is the derivation of all similar types	Fig. 3.48
	4.4 Pond		As above, but water issuing from smaller and shallow basin or doline, often representing an aquifer "eye." Such a type is the cenotes at Yucatan, Mexico if they generate flow (not stagnant water)	See Fig. 3.41k

(continued)

Table 3.3 (continued)

Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
5. Periodicity	5.1 Constant	Perennial, steady	Permanently discharging spring, despite flow variations (to prevent mixture of spring with stable discharge)	
	5.2 Intermittent	Episodic	Temporary activating spring, discharging usually during and immediately after intensive rains, floods and snow melting	See Fig. 3.411
	5.3 Periodic	Rhythmic, ebb and flow	Specific hydraulic mechanism. The pressure and water level in the connected aquifer's siphon vary causing alternate discharge and break events	 Mechanism of ebb-and-flow occurrence after Radovanović (1897). Discharge activates when water reaches maximal level in connected siphon

(continued)

Table 3.3 (continued)

Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
	5.4 Inactive	Fossil, paleo	Dry spring. Not functional anymore because of actual karstification phase which lowers water table. The synonyms "fossil," "paleo-" relevant only for that case. Evidence of ex-spring site might be presence of travertine deposits. Drying can also be result of anthropogenic impacts (e.g., forced pumping of nearby wells) in which case spring may recover	
	5.5 Estavelle		Partly ponor (sink point), partly spring (discharge point). Characterized by reversed discharge	
6. Discharge rate	6.1 Large	Strong	Habitually, with minimal flows (Qmin) over 100 l/s	
	6.2 Medium	Moderate	Habitually, with Qmin in range of 10–100 l/s	
	6.3 Small	Weak	Habitually, with Qmin less than 10 l/s, or even less than 1 l/s	
	6.4 Of first, second, etc. category (magnitude)		In accordance with average discharge Meinzer (1923b) classified springs in eight groups: 1 > 10,000 l/s, 2.1,000–10,000 l/s, 3.100–1,000 l/s, 4. 10–100 l/s, 5. 1–10 l/s, 6. 0.1–1 l/s, 7. 0.01–0.1 l/s, and 8.0 < 0.01 l/s	
7. Discharge regime	7.1 Extremely variable		Habitually, with ratio Qmin:Qmax over 1:100	
	7.2 Variable		Habitually, with ratio Qmin:Qmax between 1:10 and 1:100	
	7.3 Stable		Habitually, with ratio Qmin:Qmax under 1:10	

(continued)

Table 3.3 (continued)

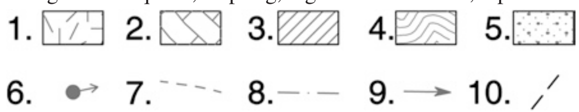
Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
8. Intake	8.1 Tapped	Captured	Springs usually with simple structure for collecting and protecting water from pollution, but not allowing changes of natural flow. Water may be used in situ (tap) or delivered by pipes for further centralizing water supply (utility)	
	8.2 Non-tapped	Non-captured	Natural discharging point without any intervention around	
	8.3 Regulated flow	Artificially arranged discharge	Complex water tapping structure enables the maintenance of or increase in the flow (storage box, pumps, wells, etc.)	
9. Water quality	9.1 Freshwater	Cold	Habitually, low-mineralized water (TDS less than 1,000 ppm) and with temperature below 20 °C	
	9.2 Mineral	Acidic, sulfuric, gas-lift, etc.	Habitually, water with mineral substances (TDS more than 1,000 ppm). Derived synonyms depend on prevailing ions, specific components or discharge mechanism	
	9.3 Thermal	Lukewarm, subthermal	Habitually, with water temperature between 20–100 °C, but synonyms are relevant for temperatures between 20–30 °C	
	9.4 Hot (water)	Hyperthermal	Habitually, with water temperature over 100 °C	

(continued)

Table 3.3 (continued)

Criterion	Type of springs	Synonyms/ other names	Short explanation	Sketch/section
10. Utilization	10.1 Drinking water supply		Satisfying potable water criteria and standards	
	10.2 Bottling water		Satisfying potable water criteria and standards. Not additionally purifying	
	10.3 Technical water supply		Non-potable springwater	
	10.4 Medicinal purpose		Used for balneotherapy or for recreation. In the former case, water contains active medicinal substances	
	10.5 Irrigation		Non-potable water but requires adequate quality to be low or slightly mineralized. The springwater often includes animals watering	
	10.6 Hydropower generation		Hydraulic head utilized for hydropower production (usually small plants)	
	10.7 Heating/ cooling		Water utilized directly (hot) or with support of heat pumps (low enthalpy resources)	

Legend for sketches/sections: *1* karstic aquifer massive rocks, *2* karstic aquifer stratified rocks, *3* impervious rocks, *4* impervious folded rocks, *5* sand, intergranular aquifer, *6* spring, *7* groundwater table, *8* potentiometric line (hydraulic head), *9* direction of flow, and *10* fault



The unavoidable mixture of criteria makes it not easy task to group or classify the springs: The same spring can be characterized from several angles, and one spring may also change its behavior depending on the water season (high/low waters).

Krešić (2010) also considered *seep* as a kind of discharge which cannot be visually observed but could be indicated by the wet soil. *Seepage spring* is a term also often used for such drainage. Diffuse flows are relatively rare in karst since concentrated flows prevail. However, such occurrences exist in wetlands, impound karst, or follow the discharge of secondary springs.

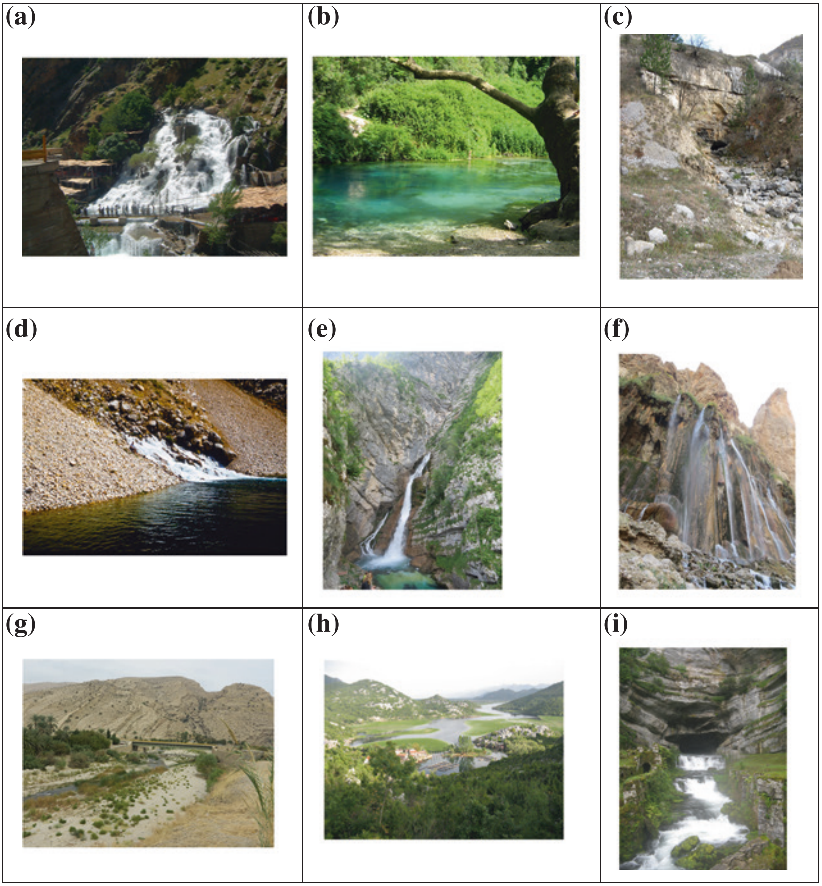


Fig. 3.41 **a** Gravity spring Bekhal (north Iraq), **b** ascending spring Syri Kalter ("Blue eye," south Albania), **c** overflow outlet of Iskretz Spring (Svoge, Bulgaria), **d** contact Istok spring (Triassic limestone–Jurassic ophiolites) (Metohija, Serbia), **e** fault gravity spring of Sava River (Bohinj, Slovenia), **f** hanging Margoon waterfall spring (Shiraz, Iran), **g** anticline plunge—Soosan spring (Kazeroon, Iran), **h** impounded spring—sublacustrine (vrulja) Karuč spring (Skadar Lake, Montenegro), **i** cave spring—Le Loue (Jura Mt. France), **j** siphonal spring of Cetina River (south Croatia), **k** pond spring—"eye" (Laas Caanood, Somalia), **l** intermittent spring (Dokan, Iraq; courtesy A.Holm), **m** small, stable and locally tapped spring Stapari (west Serbia), **n** mineral sulfuric spring Awa Spi ("white water," Sangaw, Iraq), and **o** thermal lake spring Heviz (Hungary)

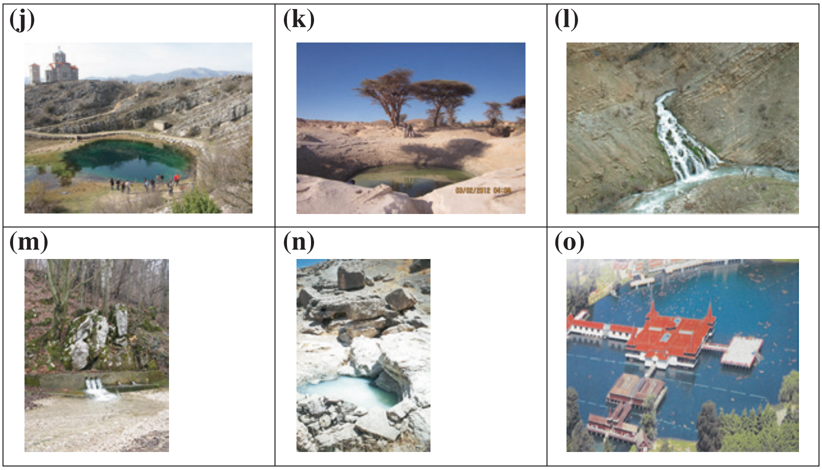


Fig. 3.41 continued

In accordance with their supposed origin Ford and Williams (2007) classify springs in terms of: (1) *emergence* (no evidence of origin), (2) *resurgence* (re-emergence of a known swallet stream, i.e., re-appearance of sinking river), and (3) *exsurgence* (autogenic seepage water).

In reference to their exact position against the shoreline, springs are *terrestrial* (*continental*) or *coastal*.

The following Fig. 3.41 illustrates some of the above-defined types of springs, but also springs which are well known, beautiful, or specific in terms of some of their properties.

Box 3.13

The complexity of the application of different classification criteria is demonstrated at Sopot Spring in Boka Kotorska Bay, Montenegro.
Sopot Spring is located at the shoreline, some 50 m from the sea, and 2 km from the city of Risan. The upper outlet is the cave 20 m above sea level which is hydrologically active only during the periods of intensive rains at the Orjen Mountain and "Stone Sea" area above Risan, which is characterized by the highest precipitation rate in all of Europe (3,000–5,000 mm/year in average). Rainy water quickly infiltrates into the highly karstified massive Cretaceous limestones where in an area of 8 km^2 more than 300 vertical shafts are registered (Milanović 2006). Depending on the saturation level, but commonly after 2–3 days, the cave as the outlet of an upper aquifer is activated, producing enormous discharges. This periodical discharge is highly impressive, and the spring cave functions with a discharge of over 150 m^3/s, one of the world's largest (Stevanović et al. 2010). Some estimates even indicate 200 m^3/s, but precise measurements are extremely difficult due

to a very steep cliff. The water then flows a very short distance from the upper cave and then with a crashing noise falls to the sea over a cascade around 20 m high (Fig. 3.42).

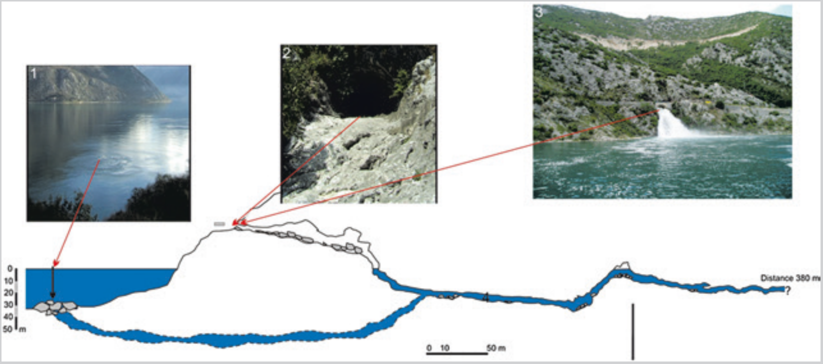


Fig. 3.42 Sopot Spring discharge mechanism (Boka Kotorska bay, Montenegro; photographs Milanović S, reprinted from Stevanović et al. 2010 with permission)

The discharge zone is predisposed by the contact of Cretaceous limestones with Eocene flysch sediments which belong to the regional Budva–Cukali tectonic zone. Direct contact is under the sea and not visible, and groundwater permanently discharges along that contact. The diving exploration located main discharge points at depths of 28 and 36 m (Milanović 2007). Research at Sopot was conducted as part of a Yugoslav-French project that lasted for 3 years. In total, 380 m of flooded canals were examined. The system is permanently discharging; during the summer season, only the two submarine springs (locally "vrulja") are active and they drain the Sopot aquifer system.

In accordance with the above site description and the proposed criteria, the Sopot Spring can be classified as follows:

- Dominantly gravity spring (but with lower ascending, and upper overflowing channels),
- Contact spring (partly impounded),
- Primary spring,
- Cave spring (with siphons at depth),
- Constant spring,
- Large spring (to moderate during the low-water seasons),
- Extremely variable spring,
- Non-tapped spring,
- Fresh water spring (with salty intrusion during the low-water seasons),
- Non-utilized spring.

It is clear that classification of such a complex discharge system may be problematic even under the same and clearly defined criterion.

The water table fluctuations and variation of pressure in an aquifer system depend on many factors such as the previously described recharge intensity, permeability, actual saturation (storage), and hydraulic head as the difference between input (recharge) and output (discharge) points. But the spring type and aquifer drainage regime are, almost by definition, closely related: The ascending springs have a more stable regime, while variation in the greater discharges characterizes gravity springs. The regime of springs which discharge from confined aquifers, such as depressional, fault, or thermal springs, is even more stable as a result of small pressure variations or limited orifices.

The spring hydrograph is thus the result of various processes which take place on the land surface or inside the aquifer system. When the main recharge is from sunken stream waters, the spring and stream hydrographs could be of similar shape. When the main recharge is from percolated rains and/or from snow melting, the differences between the maximal recharge episode and discharge peaks represent the *residence* or *travel time* necessary to estimate aquifer character. The stochastic analysis of spring hydrographs and correlative rainfall/spring flow diagram are the "books" for reading and understanding the karst aquifer behavior (Box 3.14). The method of time series, i.e., autocorrelation and cross-correlation analyses, was developed with the aim of characterizing a karst aquifer (Mangin 1984; Bonacci 1993; Krešić 2013). The correlation and cross-correlation analyses are discussed in Sects. 15.1 and 15.2.

Box 3.14

The different aquifer reaction in recharge events is presented in the case of Sarchinar Spring in north Iraq. It is one of the largest springs in the region, supplying municipal water for the Sulaimaniya city of more than 700,000 inhabitants. The outlet drains a large catchment area, some 200 km^2 of the Sarchinar–Chaq Chaq karstic system of High Folded tectonic zone (Piramagroon Mt. as a part of Zagros Mts. chain), with a dominant presence of carbonate and clastic rocks of Cretaceous age (Fig. 3.43).

Sarchinar Spring is an ascending spring issuing at the anticline plunge. The recharge of the spring is based on: (a) the diffuse infiltration of rainfall

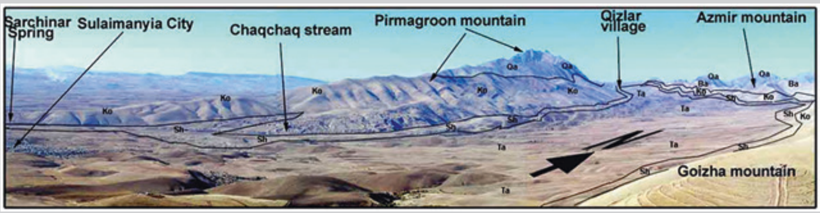


Fig. 3.43 Sarchinar–Chaq Chaq karstic system. Dominant karstic formations are Kometan (*Ko*) and Qamchuga (*Qa*), while low permeable rocks belong to Shiransh (*Sh*) and Tanjero (*Ta*) Fms

through the exposed outcrops of thickly bedded and highly fractured limestone layers; (b) the percolation of runoff and intermittent stream water of Chaq Chaq Valley through tectonically active zones (Ali et al. 2009). During the recession periods of the extremely dry years 1999 and 2000, the minimal discharge of the system was around 600 l/s. The recorded maximum during the period of 1999–2005 was 7,454 l/s (March, 2003).

The main factor which directly reflects the regime of karstic aquifers in the study area is related to the unequal distribution of recharge. The rainy season usually ends in April and no single rainfall event would occur until late September. During the rainy season, the system reacts to rainfall events with a delay of a few days representing the minimum travel time for the recharging inputs (Fig. 3.44). Meanwhile, a slow reaction could be observed during the low-water periods after approximately 1 month (rainfall at the beginning of May transferred as output at the beginning of June). Further on, during a recession period, the spring hydrograph displays a typical monotone depletion of the accumulated resources as a base flow (Stevanović and Iurkiewicz 2009).

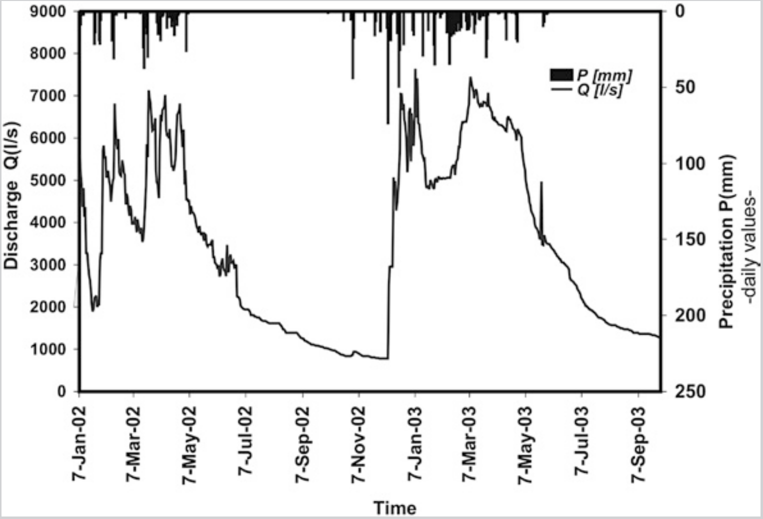


Fig. 3.44 Hydrograph of Sarchinar Spring (after Stevanović and Iurkiewicz 2009)

The regime of this spring is determined by several factors. The principal factors are arid climate and cyclic recharge variations (6 months without rainfall) on one side, and significant aquifer storage capacity and slow drainage on the other. Recession analysis and obtained coefficient α confirm a very slow drainage: It was inferred that to exhaust the dynamic resources of Sarchinar reservoir completely would theoretically require a period of several years of continuous discharge without any additional recharge (Ali et al. 2009)

Krešić (2013) presented the complex shape of hydrographs based on Jevdjević's (1956) explanations of the functioning of single hydrographs. The single hydrograph shows the transformation between the input function (precipitation) and the output function (discharge). The recharge waves may pass quickly through the system but may also simply accumulate if the deficit in stored reserves is large (after long recession). In fact, the common complex hydrograph is a result of the superpositioning of single hydrographs which correspond to separate rainfall episodes (Fig. 3.45).

Mangin (1984) and Padilla and Pulido-Bosch (1995) applied correlation and cross-spectral analysis on several karstic springs in France and in Spain and made an attempt to generalize results obtained from a single hydrograph of the spring. The parameters that can be assumed are the *response time*, the distinction between

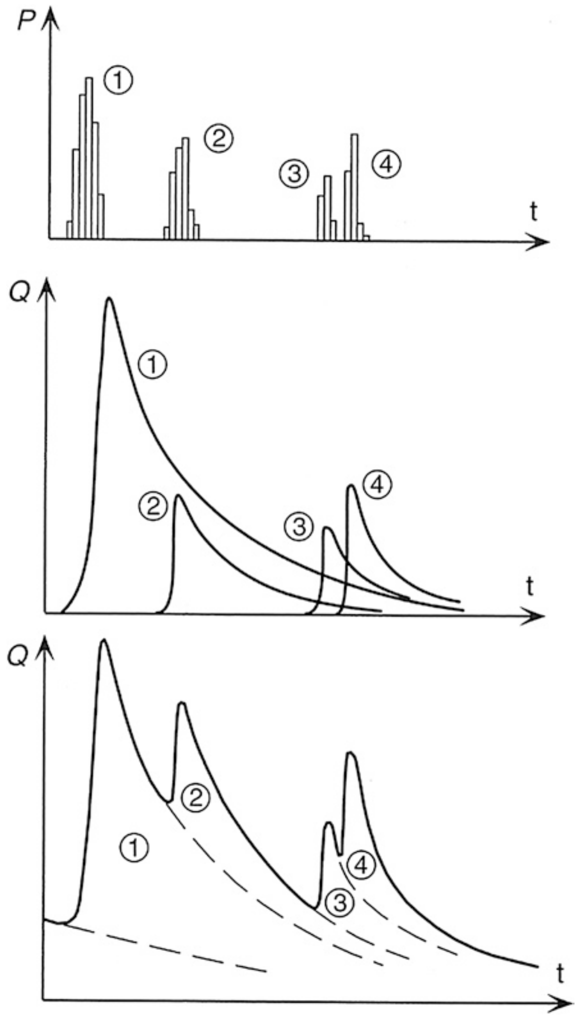


Fig. 3.45 Complex spring hydrograph resulted from several rainfall events (1–4). *P* precipitation, *Q* discharge (after Jevdjević (1956) and modified by Krešić, reprint with permission from Water in karst (2013), Mc Graw Hill Education)

quickflow, *intermediate flow* and *base flow*, and the *mean delay*. They stated that the "method offers quantifiable and objective criteria for differentiation and comparisons of karstic aquifers." Figure 3.46 shows four typical single hydrographs with various memory effects (prolongation of recharge on hydrograph shape) proposed by Mangin to be widely applied as etalons on similarly obtained hydrograph shapes.

Padilla and Pulido-Bosch (1995) additionally examined three out of four etalon springs and confirmed lag time, i.e., response of aquifer to rainfall events. The response at Aliou and Baget was immediate, while at Torcal was after 12–35 days.

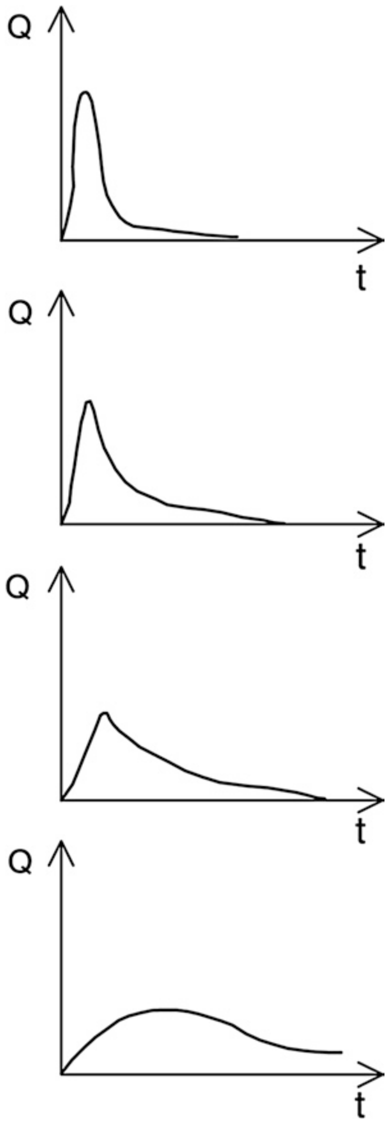


Fig. 3.46 Mangin's typical single hydrographs of four tested springs proposed as etalons. From the *top Aliou* (memory effect: reduced, up to 5 days); *Baget* (memo: moderate, 10–15 days); *Fontestorbes* (memo: large, 50–60 days); and *Torcal* (memo: significant, 70 days)

Iurkiewicz (2003) presented various single hydrographs as a unit step response function for the surveyed springs in the Banat Mts. in Romania. Figure 3.47 shows results obtained for the central compartment (Miniş–Nera zone).

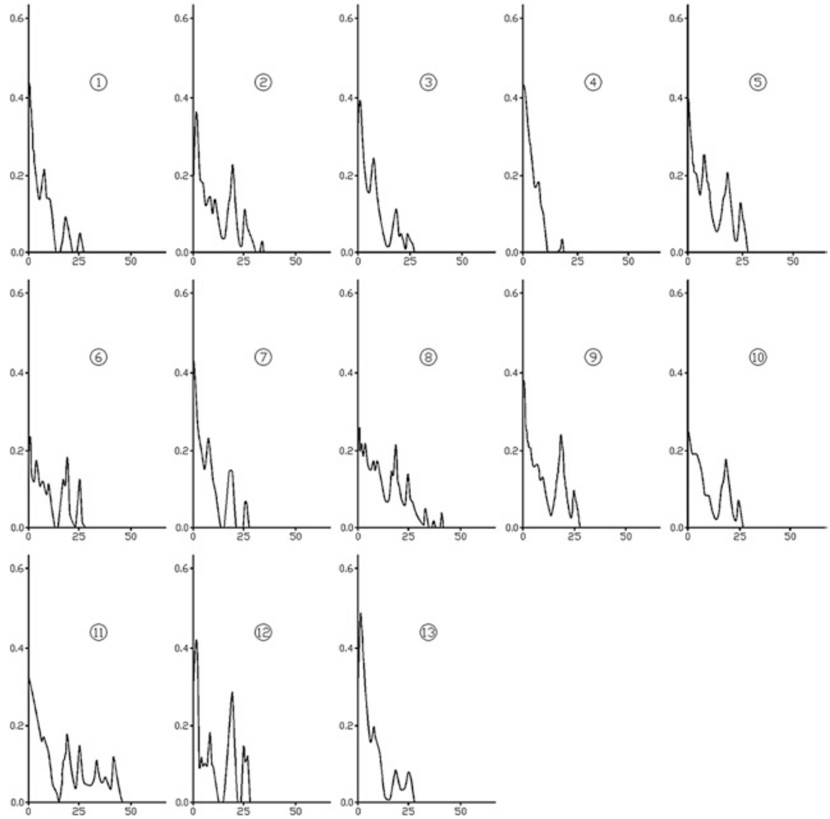


Fig. 3.47 Single hydrographs of karstic springs in Miniş—Nera zone (Romania) (after Iurkiewicz 2003)

Box 3.15

The Fontaine de Vaucluse is one of the most famous and best explored karstic springs in the world. It is located in Provance in southern France, about 30 km from the city of Avignon. The deep siphonal channels, enormous variation in discharges, and great minimal flows have always attracted researchers from all over the world. Blavoux et al. (1992) define the catchment area of the Fontaine de Vaucluse system as over 1,100 km^2. The Lower Cretaceous limestones of Urgonian facies are 1,500 m thick and highly karstified. The thickness of the unsaturated zone in the entire basin is over 800 m; four sinkholes at the plateau are more than 500 m deep, but their bottoms still do not reach the saturated zone (Blavoux et al. 1992).

The spring is a siphonal lake which is an upper outlet and not functional during the low-water seasons. The huge fallen blocks are masking the downstream part, and springwater is always discharging through this thick debris creating the Sorgue River, one of the tributaries of the Rhône. The water fluctuates in the discharge zone for about 25 m in an average hydrological year (Fig. 3.48).

Fig. 3.48 The Fontaine de Vaucluse in a low-water period in August, 2011. *Left photograph* deep siphonal channel and dry water gauges on the wall. *Right* Sorgue river flow some 300 m downstream from spring site (discharge from debris)

The Fontaine de Vaucluse is thus a lake spring with a deep siphonal channel used as *locus typicus* for all such springs worldwide. It is surveyed by many speleo-divers and robots (remotely operated underwater vehicles, sonors) that also were constructed for exploring deep channels. The Vaucluse's museum "Le Monde Souterrain" (The Underground World) provides information on the long history of these surveys starting with Ottonelli in 1878, but also undertaken by the famous Cousteau (in 1944, 1955), Touloumdijan (in 1980s), and many others. The deepest point of the siphon at −308 m was reached by the machine the "spelenaute."

The average spring discharge is around 20 m^3/s, making that spring the largest in France. The spring discharge has been recorded since 1878 making this spring one of the best explored in the world in terms of drainage regime. Minimal discharge is 3.7 m^3/s, while the maximum exceeds 100 m^3/s (Blavoux et al. 1992).

Cognard-Plancq et al. (2006) confirmed quick system responses to rainfall in comparison with the large recharge area: The peak of the hydrograph occurred 24–72 h after the rainfall events. The springwater level and discharge depletion are slow, which can be explained by the existence of the large storage capacity of the aquifer. Bonacci (2007) also studied the discharge regime for 127 years (1878–2004) and identified a generally decreasing trend of aquifer discharge equal to 0.0468 m^3/s per year (Fig. 3.49).

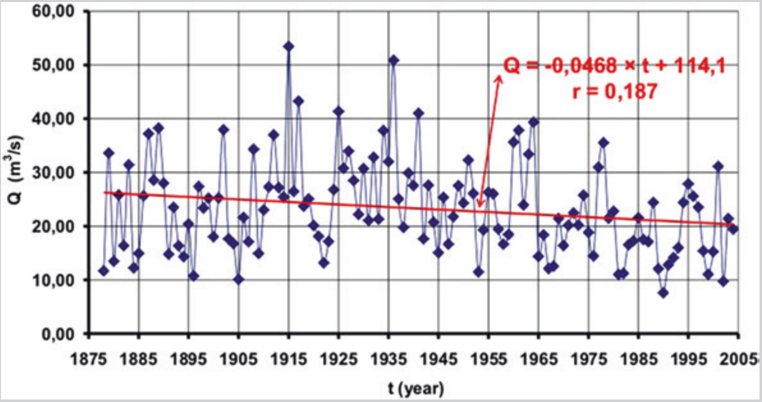


Fig. 3.49 Time data series of average annual discharges (Q) at the Fontaine de Vaucluse with trend line for the period 1878–2004 (after Bonacci 2007, printed with permission)

That trend, although not significant, is not easy to explain because during the same period, the annual rainfall in the catchment has an increasing trend. Along with a question on the accuracy of the measured spring flows and rainfalls, or problematic delineation of the catchment, Bonacci (2007) noticed that the weak relationship between runoff and rainfall might be the result of some others factors such as air temperature, groundwater level, interannual rainfall distribution, changes of catchment area during the time, preceding soil wetness, anthropological influences, and climate change.

Stochastic methods and an established relationship between input–output signals enable not only the time data series to be refilled by missing (unmeasured) data but also forecast discharges under different climatic scenarios. For instance, prior to a technical decision to tap or not tap a karstic spring, it is possible and even advisable to estimate karstic aquifer drainage behavior under long drought episodes (with limited or no recharge). Figure 3.50 shows hydrographs and results obtained from the created stochastic model of Veliko Vrelo Spring in the Carpathian karst of Serbia. The established rainfall–discharge function for

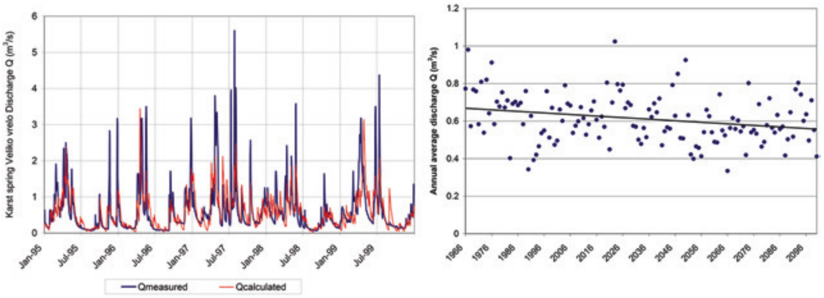


Fig. 3.50 Measured and simulated discharges (*Q*) of Veliko Vrelo Spring (*left*, after Stevanović et al. 2010, printed with permission) and forecasted annual mean discharges of the same spring including general trend up to end of 21 Ct (from Stevanović et al. 2012a)

daily values is used as one of the parameters for long-term prediction of spring discharges (see also Box 15.2.5 in Sect. 15.2). The time series of annual mean discharges of Veliko Vrelo are obtained by using the bias-corrected regional climatic model E-obs for precipitations and air temperatures for the period until the year 2100 (Stevanović et al. 2012a). The result shows a depletion of groundwater reserves by around 15–20 % for the predicted climatic scenario.

The karstic springs are widely utilized as a source of drinking water supply. It is known that some 20 % or even a few percent more of the global population largely depends on karstic groundwater (Ford and Williams 2007), but there are no statistics yet as to how much of this water is provided directly from the springs and how much from other structures. Discussion on actual and possible problems when natural spring flow is not regulated is provided in Sect. 15.5.

Many countries utilize karstic springs simply because there are no other alternatives, but in many other countries, awareness of their importance and the good water quality they provide is a principal factor for such a decision. The karstic aquifers have a significant proportion of the water supply in the following regions: southeastern Europe (Alps and Carpathians, Box 3.16), the Mediterranean basin, the Near East and Middle East, the Arabian Peninsula and Horn of Africa, southeastern Asia, North Africa, the Caribbean basin and Central America, and the southern part of the USA.

Box 3.16

Drinking water supply for about half of the population or four million citizens in Austria is effectuating from karstic aquifers (Zötl 1974). Vienna has the largest water supply system based on karst waters. The system consists of two major gravity pipelines, 130 and 200 km long (Drennig 1973). The first Viennese mountain spring pipeline was completed in 1873 by tapping water from the Kaiserbrunn (Fig. 3.51) and other springs issuing from Schneeberg, Rax, and Schneealpe Mountains, while the second pipeline was completed in 1910 by tapping the Kläffer and other smaller springs discharging from Hochschwab Mountain (Styria). The system was established as an alternative to Danube alluvial waters which often caused hydric epidemic. The total catchment is around 600 km^2. The long concrete tunnels and channels provide an average yield of 4.5 m^3/s, for some 1.7 million inhabitants. These two springwater supply pipelines which meet around 95–97 % of the amount of water required for the municipality of Vienna (on average 140×10^6 m^3/year). The water is of excellent quality. Generally, no treatment is applied; only chlorination is required, primarily for cleaning the distribution pipes.

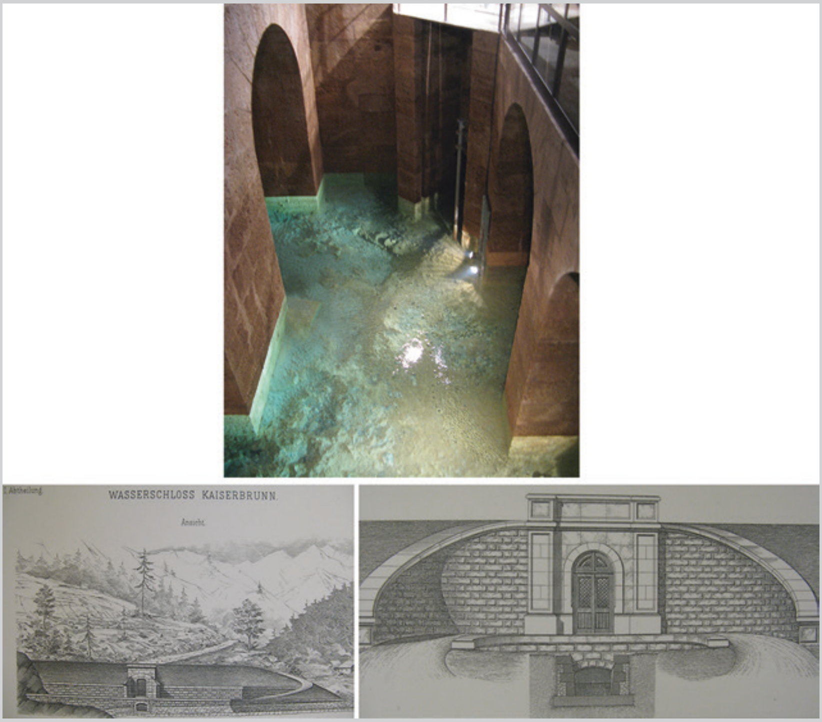


Fig. 3.51 Kaiserbrunn's old design sketches (photographs made by courtesy of the Kaiserbrunn Museum, Vienna) and spring's intake (*inside view*)

Natural drainage of aquifers through springs can therefore cover water demands on a wide scale: from the supply to multi-millions of towns at the regional level, to the very local level where the supply to just one or several houses is concerned. Although the latter is not a big problem in terms of amount of water, for the big consumer a very large aquifer and spring discharge are required. The population

Fig. 3.52 Camels around one of the rare freshwater springs draining Taalex Fm. in Puntland province of Somalia

growth and increased demands have caused many large cities to substitute or enhance their primary water system based on springwater, with surface waters or groundwater from other aquifers (Stevanović 2010b).

Table 3.3 and criterion *Utilization* present a wide range of uses of karstic waters. In arid regions in the Near East and Middle East, it is, for instance, very common to tap karstic springs and to construct gravity channels for irrigating arable land. Use of pipes instead of channels significantly increases efficiency in water provision because losses from channels are usually very high due to seepage or evaporation resulting from systems' openings. Springwater is also widely used for watering animals, and fresh water of good quality provides security for animal health and growth. Thus, in a rural environment, it is common to see a large numbers of animals occupying the springs or the ponds and swamps formed nearby (Fig. 3.52).

The use of karstic waters in hydropower generation by utilizing high hydraulic head is limited mostly to the Alps (Austria, Switzerland), while thermal properties of karstic waters and springs are utilized elsewhere. Finally, the number of karstic springs utilized in the world's water bottling industry which runs an annual revenue of around $13 billion is very large and karstic aquifers probably lead the list of aquifers where such sources originate.

The conflict of interest in utilizing karstic waters is present especially in the undeveloped world. This is not very different from conflicts related to any other aquifer or surface water, but the recognized precious water quality of karstic sources may sometimes be an additional factor for disagreement.

3.6 Quality of Karst Groundwater

The dissolution of rocks and the duration of direct water–rock contact results in variable groundwater quality at discharge points. The mineral components of karst waters depend upon the composition of the rocks through which water is percolating: Hydrocarbonate (HCO_3)–calcium (Ca) type of waters is created from the dissolution of calcium carbonate which is a dominant type of water in limestone, (see Box 3.17), while the hydrocarbonate (HCO_3)–magnesium (Mg) type of groundwater is present to a lesser extent, and is regularly connected to dolomitic rocks.

Langmuir (1984) listed processes which control and influence the quality of groundwater before it reaches the spring site or well head. These processes are as follows:

1. The composition of the infiltrated atmospheric precipitation;
2. Evapotranspiration losses from groundwater recharge and shallow groundwaters;
3. The acidity and degree of undersaturation of groundwater recharge;
4. The availability and solubility of carbonate and associated rocks, including halite, gypsum, and anhydrite;
5. Rates of solution of the rocks and contact time;
6. Hydrological processes such as dilution by fresh water recharge and mixing of dissimilar groundwater;
7. Anthropogenic processes, including groundwater pollution by wastes and leachates from solid wastes.

Calcium carbonate is highly soluble when carbon dioxide is present, and its concentration in water ranges from 200 to 300 mg/l under normal conditions. The sodium chloride and calcium sulfate waters, which result from the dissolution of

Table 3.4 Chemical composition and general water quality of some karstic aquifers (after Stevanović and Papić 2008, modified)

Karstic aquifer	Lithology	Minerals	Chemical composition	Expected quality	Remarks (possible problems for drinking water quality)
Carbonate	Limestones	Calcite	Hydrocarbonates, calcium	Low mineralization (0.3 g/l), pH 7–7.5	Problem: turbidity, microbiology, nitrates
	Dolomites	Dolomite	Hydrocarbonates, calcium, magnesium	Medium mineralization (0.5 g/l), pH ~ 8	Problem: hardness, iron
	Marbles	Calcite (dolomite, quartz)	Hydrocarbonates, calcium, (magnesium, silica, iron)	Low mineralization, pH ~ 8, iron	Problem: hardness, iron turbidity, microbiology
Evaporite	Gypsum	Gyps	Sulfates, calcium	Higher mineralization (2–3 g/l), pH ~ 6	Problem: bitter odor
	Halite	Halite	Chloride, sodium (sulfates, calcium)	Very high mineralization, salty water	Problem: salty, bitter odor

halite and gypsum, are undesirable from a water supply perspective due to changes in the organoleptic properties of water (Stevanović and Papić 2008). The issue of rock dissolution is discussed in Sect. 2.3.

The typical groundwater quality of water issuing from different karstic aquifers is presented in Table 3.4.

Groundwater quality can also be the result of a cation exchange process, which is a frequent occurrence in limestones whose fissures may include clay minerals; as such, an initial hydrocarbonate–calcium type of water may be replaced by a hydrocarbonate–sodium composition. This is a typical example of a natural softening of "hard" calcium waters by "soft" sodium waters. Herak et al. (1981) stated that even small lenses or intercalations may strongly influence the chemical properties of the water; thus, it is almost impossible to give a common composition of waters in karst.

Box 3.17

The typical HCO_3–Ca or HCO_3–Ca/Mg type of water characterizes the carbonate karst of the Carpathian and Dinaric mountain arches in SE Europe. The chemical composition of these waters reflects entirely the conditions of their formation, the intensive water exchange, and the rapid filtration.

- Around 300 analyzed samples from the Carpathian karst of Serbia confirmed low mineralization of waters (total dissolved solids TDS of 0.2–0.4 g/l, only 10 % samples below, and 6 % above these values) with a prevailing content of HCO_3 in the anion and of Ca in the cation composition. The average content of HCO_3 is 87 % mval and of Ca is 75 % mval, whereas the ions Mg, Na, K, SO_4, Cl, NO_3, usually occur inferiorly. The ions NH_4, Fe, Mn, as well as microelements, are usually absent in results of conventional laboratory analyses, and if they do occur, it is always within the limits of the drinking water quality standards.
- Based on chemical analyses of around 180 samples taken from karstic aquifers in Dinaric karst, Petrik (1976) concluded that $CaCO_3$ hardness was between 100–250 mg/l for 50 % of all samples. In continental Dinaric karst, ion Cl is usually present in groundwater with a concentration less than 10 mg/l, but along the coast and in the islands, it may significantly increase (over 5,000 mg/l) due to sea water intrusion.
- Dynamic karstic water regime may be very problematic concerning low discharges in recession periods, but during the floods, we often witness another problem of high turbidity and considerable change in natural water quality. Water temperature in unconfined aquifers regularly does not exceed annual amplitude of 5–6 °C. For instance, those are usual values for Fontaine de Vaucluse (maxima in late summer and minima in late winter), while in Carpathian karst, the average amplitude of groundwater temperature regularly does not exceed 3 °C. The water mineralization throughout the hydrological year rarely varies higher than 0.1 g/l.

Evaporitic rocks bind more soluble characterize water with higher salinity and increased content of Cl and SO_4 anions.

- Figure 3.53 shows diagrams of low-mineralized springwaters issuing from carbonate Cretaceous and Paleogene aquifers in central part of northern Iraq (Qamchuga, Dokan Fms.) and mineralized springwaters from Neogene evaporitic aquifers (Fars Fm. Garmian area, N Iraq).

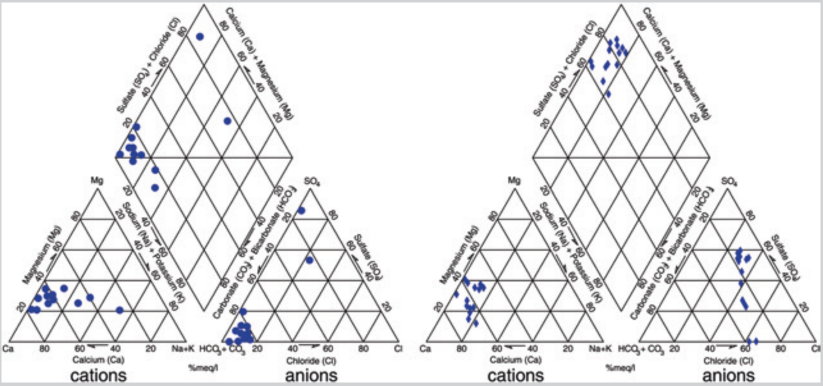


Fig. 3.53 Piper's diagrams of low-mineralized karstic springwater from carbonate Cretaceous and Paleogene aquifers (*left*) and mineralized springwater from Neogene evaporitic aquifers (*right*) in northern Iraq

- In the Somali provinces of Somaliland and Puntland (the Horn of Africa), both carbonate and evaporite karstic aquifers are present. In terms of geological formations, water from carbonate aquifer of the Jurassic age is the best quality (Stevanović et al. 2012b). Water mineralization is moderate and the electro conductivity (EC) is commonly around 600 μS/cm. The bicarbonate type of water prevails, but concentration of sulfate and chloride might be occasionally slightly higher in some samples. In contrast, in evaporite karst of Eocene Taalex Fm (Fig. 3.54), EC is very high, ranging from minimal 890–7,270 μs/cm. Evaporites are generated during the precipitation from an over-saturated brine solution which is usually highly soluble where concentrations rapidly increase, even if the portion of evaporitic material is relatively small. When recharging fresh water comes into contact with a thin layer of anhydrite or gypsum, the EC will rapidly increase (usually beyond the potable limit). Water is often highly sulfatic, with calcium as the predominant cation. SO_4 is in the range 125 up to 3,100 mg/l, with an average concentration of 1,300 mg/l.

Fig. 3.54 Fissured gypsum of Taalex Fm. (Somaliland)

Analyzing rock-water interactions, Younger (2007) concluded that due to low velocities of groundwater in saturated zones, there is normally sufficient time for even slow geochemical reactions to alter water chemistry considerably. Dissolution of calcite under laboratory conditions requires no more than 24 h to approach equilibrium; considering the years or even centuries of calcite–water contact in saturated zones that equilibrium is more standard than exceptional in limestone aquifers. Therefore, even brief contact between percolated water and karstic rocks may sometimes significantly change the water quality (Box 3.18).

Box 3.18

Investigating causes of an accidental water leakage which appeared in several houses located on the slope of a hill formed from Miocene vuggy limestone in the Belgrade suburb Rakovica (Serbia), water samples were taken from the waterworks' pipeline on the top of the hill, and from a small cave 40 m beneath the top of the hill (Fig. 3.55). The seepage sites followed a thin marly horizon interstratified as lower permeable layer between upper and lower limestone horizons. The chemical composition of the water from the pipeline and underneath the cave was almost equal, confirming the direct infiltration and fast percolation in Miocene aquifer of artificially treated chlorinated water. The electrical conductivity of 655 μg/l and trihalomethanes as a residual product of water chlorination of 7.2 μg/l indicate that seepage is probably from the crashed pipeline. Although not far from the cave, the quality of water which further filtrated through limestones of vuggy porosity quickly changed and adapted to become similar to natural springwaters

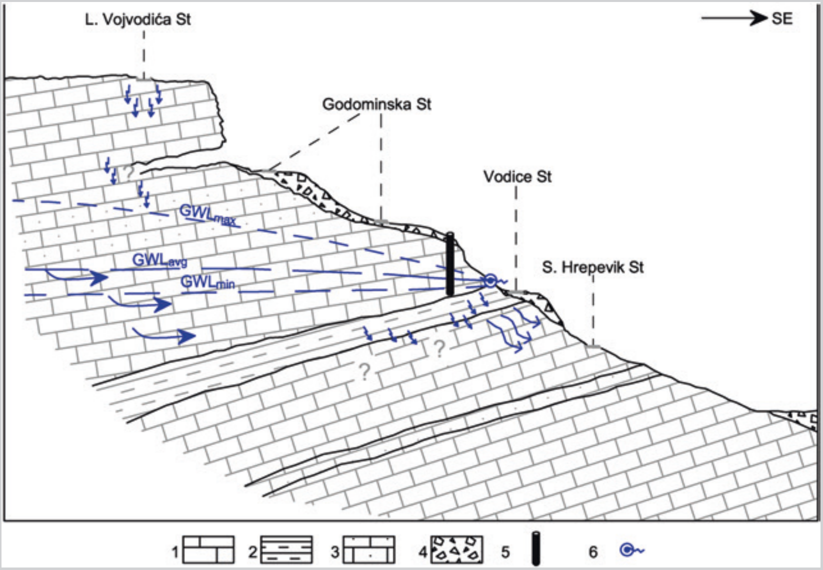


Fig. 3.55 Cross section of Miocene limestone aquifer in accident area in Rakovica, Belgrade. *Small arrows* indicate leakage and percolation toward houses in Vodice Street. *Legend 1* Miocene limestones, *2* miocene marls, impermeable barrier to groundwater flow, *3* miocene sandy limestones, *4* debris, *5* dug well, *6* seepage sites. *GWL* groundwater levels (max–min)

which drain Miocene aquifer. The seepage water sampled in the basements of houses had an increased salinity, almost double that in the pipeline (EC of 1,105 μg/l), while the content of trihalomethanes was reduced to 2.2 μg/l. Although the latter value was still greater than the natural one, the reason for this accidental leakage was clearly recognized as damage on the pipeline was found and then repaired. As an additional remedial measure, the groundwater table was depleted by forced over-pumping of a nearby dug well and leakage was soon stopped. In conclusion, even though the travel time through very porous media was very short, the water quality had significantly changed.

It is almost a rule that groundwater in open karst structures is low mineralized which is a result of the intensive water exchange and rapid filtration. In deeper parts of the aquifer, slower filtration results in an increase in mineralization. This variation is often minimal, but nevertheless indicates a certain differentiation that may be important under specific circumstances (e.g., when pollution is involved). Similarly, when the conventional classification into gravity and ascending springs is concerned, in principle the latter are easier to protect because they drain "lower" and slower circulation zones. When a specific type of "combined" springs with a drainage system consisting of gravity channels and ascending siphons in the same discharge zone is indicated, the water quality of various channels should be monitored because irrespective of the “unity” of a karst aquifer, local variations in water quality are very probable.

The best waters in terms of quality are generally those from the deepest parts of an aquifer, particularly water near or below the base level of erosion or discharging underground into laterally connected aquifers. In these parts of a karst aquifer, the self-purification of water is the highest, and the quality, regardless of possible slightly increased mineralization, is always the best. This of course excludes the very deep aquifer zones where the water exchange might be very slow, while the temperature might also increase as a result of a natural or possibly an anomalous geothermal grade.

The above facts led us to consider a hypothetical “bedding” character of karst groundwater and separation of hydrochemical zones:

1. a zone of fast water exchange and propagation which corresponds to the highest levels of respective karst channels, the least favorable hydrochemical and bacteriological conditions for protection;
2. a zone of slow water exchange with dominantly horizontal or siphonal circulation, including the subsurface outflow area (always of the best qualitative properties), and, finally,
3. a deep zone of retarded water exchange, often unsuitable for water supply because of the increased temperature, mineral content or presence of specific micro constituents.

In reference to the above, Langmuir (1984) noticed that waters extracted from wells often have higher TDS than adjacent springwater in the same aquifer. This is because springwaters generally discharge from zones of enlarged porosity, whereas randomly sited wells most often draw water from less permeable zones.

Isotopic analyses may significantly contribute to assessing groundwater origin, contact time, recharge conditions, and consequently even the catchment size. Most of the analyses concerns stable environmental isotopes such as ^{2}H (deuterium—D), ^{16}O, ^{18}O (oxygen 16 and oxygen 18), ^{13}C (carbon 13), and radioactive man-made isotopes (as a result of nuclear explosions) such as ^{3}H (tritium) or ^{14}C (carbon 14).

Due to their openings and fast groundwater circulation, karst aquifers have very low attenuation capacities and are extremely vulnerable to pollution. This issue is discussed in several contributions in Chap. 17. In addition to easy infiltration of the contaminants, the results of many tracing experiments indicate that under convenient conditions, the pollution may migrate very fast, as far as 1 km of rectilinear distance in as little as 24 h. Active hydraulic connections between the ground and surface waters, where the harmful components may be carried from great distances including non-karst terrains and infiltrated into the narrowest spring zone, have a particular significance in this regard.

Therefore, in the case of carbonate karst, the quality of natural karst waters is excellent almost by definition: It is confirmed in many places worldwide that water issuing from unpopulated catchment areas on mountain massifs is sanitary and pure while only exceptionally could there be a small amount of bacteria present, but if pollution sources are present in the catchment of an unconfined karstic aquifer, then severe hazards follow (Box 3.19).

Box 3.19

The water supply of the town Shiraz in central–south Iran was initially based on the tapping of karstic springs which drain the Asmari limestone formation (Oligocene–Lower Miocene age). The city is located in a flat area representing the core of syncline which consists of impervious marly sediments of Mishan Fm. (Middle Miocene). The Mishan's sediments are very thick, reaching some 800 m. The basin is surrounded by high hills consisting predominantly of Asmari limestones, and along the marginal parts and in contact with Mishan Fm. (and adjacent underlying Gachsaran Fm.), many large springs are located making this a suitable site for the establishment of this city which further grew and became the strategic center of the Persian Empire.

Figure 3.56 presents the three stages of evolution of karstic aquifer in accordance with the concrete intensity of water utilization. The normal

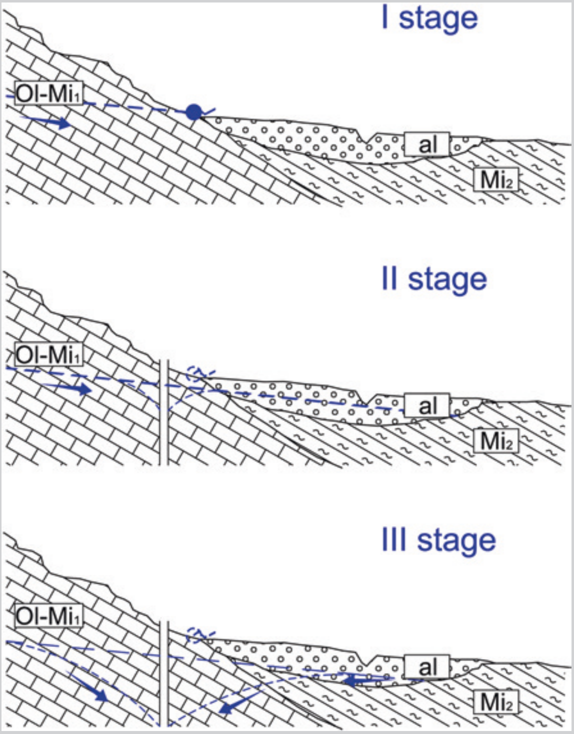


Fig. 3.56 Evolution of the water table during three stages of extraction of groundwater for water supply of the city of Shiraz in Iran. *Legend* Ol-Mi_1 Asmari limestones—karstic aquifer, Mi_2 Gachsaran and Mishan Fms.—impervious rocks, al—alluvial sands, gravels, clay—intergranular aquifer

functioning and natural discharging of the spring marks stage I. Along with city growth (today nearly 1.5 million inhabitants) and the pumping of drilled wells also located along the Asmari–Mishan contact, the springs started to dry periodically, and afterward, with further intensification of groundwater extraction, the springs completely dried out (stage II). The further aquifer development resulted in significant depletion of the groundwater table and the reverse infiltration of groundwater from the alluvium of a temporary stream which passes through the city center (stage III). Currently, a major problem is the pollution of alluvial and, consequently, of karstic aquifers by nitrates resulting from infiltrated sewage and irrigation waters.
(This situation is explained in personal communication with Dr Ezzat Raeisi from Shiraz University, Iran).

The direct impact of pollutants on the karstic aquifer might in some cases be reduced and even minimized. Figure 3.57 shows several case examples of the functioning of a single karstic channel or siphon and its role in purifying or amortizing the impurity of the water flow through it. The siphon could be pressurized with a potentiometric line above the channel (cases a, c, e) or with gravity-free flow (unconfined, cases b, d, f). Additionally, the siphon may be totally empty of any sediments (a, b), or partly filled with sediments (c, d). Finally, the end of a channel could be sealed with some fine sediment which belongs to laterally connected aquifer of intergranular porosity (e, f). Each of these options results in a specific water quality. There are also many factors which influence the duration of rock–water contact such as potentiometric pressure, channel length and inclination, permeability of sediments' plug, and water viscosity. However, the cases presented can be used as a starting point for estimating whether the attenuation effect on contaminated water could be smaller or larger.

Figure 3.57 contains diagrams showing the expected reduction of bacterial concentration in water samples for each of the presented cases. Experimental determinations of the life cycle of different microorganisms in limestones are very important. Depending on the conditions, the shortest life of *Escherichia coli* is 70–210 days, which proves how old pollution of this type is. At low temperatures (4–8 °C), life cycles vary from 40 days for *Salmonella* (at contamination rates to 10^5 microbes/l of water) to 120 days for enterococci (at contamination rates to 10^8 microbes/l of water) (Gavich 1985). Therefore, slower water movement through channels and longer rock–water contact proves the reduction of bacteria in the system. The worst scenario for water quality presented on Fig. 3.57 is first presented case indicated by (a) an empty and pressurized channel. From the point of view of water purification, every additional case presented is better, while significant improvement of water quality and even elimination of bacteria may be expected when an unconfined siphonal channel is laterally connected to fine porous sediments (case f). The last case can be similar to the functioning

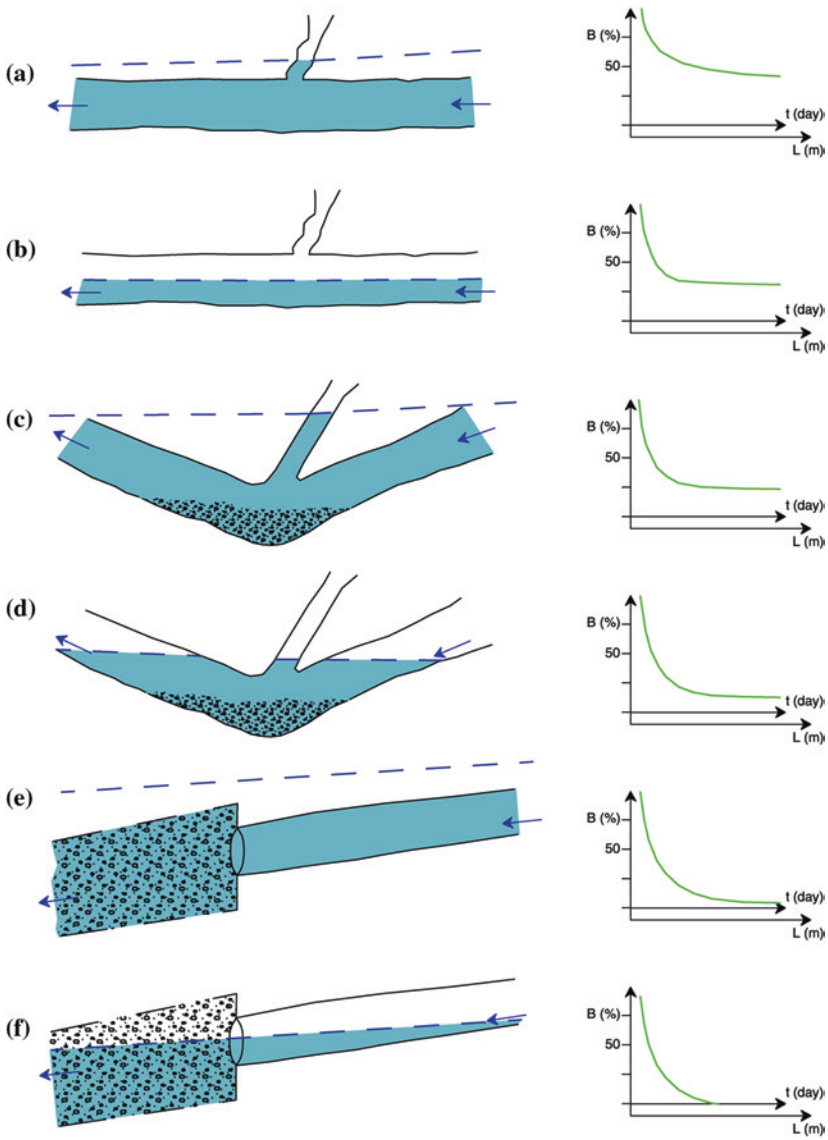


Fig. 3.57 The karstic drainage channel: **a** pressurized and empty, **b** unconfined, empty, **c** pressurized, knee shaped, partially filled with sediments, **d** unconfined, knee shaped, partially filled with sediments, **e** pressurized, with adjacent porous aquifer, **f** unconfined, with adjacent porous aquifer. The diagrams on the *right* show a reduction of bacteria (*B*) in the percentage based on residence time (*t*) and the length of the channel (*L*), i.e., distance between input and output points

of a secondary spring type. There are many such springs worldwide located in mountain foothills. Probably, the most famous and well explored are those issuing from Alps mountain chain. Some of them are located at the northern margin of Le Mans Lake in France, in proximity of Evian mineral water source. These springs discharge from fine lacustrine–glacial sediments to which water partly arrives from the adjacent high-Alpine karstic aquifer with catchment altitudes reaching 2,000 m.

Finally, after detailed discussion of karstic aquifer properties, we can repeat in conclusion that each karstic aquifer is an individual case. Nonetheless, we may still try to highlight some general outlines of the character of karst aquifer:

Accessible. Not everywhere and not always, but still reachable even in deep submerged siphons, if not by humans, then by instruments of navigation. It is just a matter of time before the new technology (scanners, acoustic log, floating sensors, nanotechnology or something else) will enable complete tracing of main and secondary channels and complete recognition of groundwater paths.

Variable. Karstic aquifer is a non-homogenous and anisotropic system changing its properties from place to place. It is a fact that for a minimal distance, one may find an extremely permeable cavity and a totally compact impermeable block. Similar to spatial variability, the karst aquifer system is also time variable. The springs which discharge enormous water during the flood periods often become dry during the long recession periods.

Unpredictable. The system is dynamic with a lot of changes during a hydrological cycle. Many engineering solutions and designs have failed because of karst aquifer variability but also due to improper or insufficient research. However, it also true that prior to construction of any type of structure in karst, properly conducted investigations and an appropriate project concept can significantly reduce the risks or minimize them to acceptable levels.

Precious. No other aquifers can provide water of such high natural quality as carbonate or dolomitic karst can (Fig. 3.58). Generally, low-mineralized water with dominant HCO_3 and Ca, Mg ions is an ideal arrangement for the human

Fig. 3.58 Pristine low-mineralized water of cave spring Genal in Sierra de las Nieves (south Spain), however, characterized by rapid chemical variations (personal communication B. Andreo)

organism. In many places and utilities worldwide where pollutants are absent, chlorination is the only required treatment of water.

Vulnerable. Due to the presence of large voids and cavities as preferential paths which also cause turbulent flows, karstic groundwater can be easily polluted and pollutants can be quickly transported a long distance. Therefore, the pollution risk is much higher than in the case of other aquifers, but more problematic is the very limited attenuation capacity of karst (see Chap. 8 and Sects. 17.1–17.4).

Beautiful. As mentioned repeatedly in Chap. 2, there are so many natural wonders created in a karstic environment. Some are already protected, while others await further evaluation and decisions before they are eligible for inclusion on the long list of internationally or locally protected objects.

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