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Hydrogeology of Karstic Area

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1. Introduction

Karst is a special type of landscape that is formed by the dissolution of soluble rocks. Karst regions contain aquifers that are capable of providing large supplies of water. More than 25 percent of the world's population either lives on or obtains its water from karst aquifers. In the United States, 20 percent of the land surface is karst and 40 percent of the groundwater used for drinking comes from karst aquifers. Natural features of the landscape such as caves and springs are typical of karst regions. Karst landscapes are often spectacularly scenic areas.

2. Karst definition and different types of karst

The term karst represents terrains with complex geological features and specific hydrogeological characteristics. The karst terrains are composed of soluble rocks, including limestone, dolomite, gypsum, halite, and conglomerates. As a result of rock solubility and various geological processes operating during geological time, a number of phenomena and landscapes were formed that gave the unique, specific characteristics to the terrain defined by this term. Karst is frequently characterized by karrens, dolines (sinkholes), shafts, poljes, caves, ponors (swallowholes), caverns, estavelles, intermittent springs, submarine springs, lost rivers, dry river valleys, intermittently inundated poljes, underground river systems, denuded rocky hills, karst plains, and collapses. It is difficult to give a very concise definition of the word karst because it is the result of numerous processes that occur in various soluble rocks and under diverse geological and climatic conditions (Milanovic, 2004).

The main features of the karst system are illustrated in Figure 1. The primary division is into erosional and depositional zones. In the erosional zone there is net removal of the karst rocks, by dissolution alone and by dissolution serving as the trigger mechanism for other processes. Some redeposition of the eroded rock occurs in the zone, mostly in the form of precipitates, but this is transient. In the net deposition zone, which is chiefly offshore or on marginal (inter- and supratidal) flats, new karst rocks are created. Many of these rocks display evidence of transient episodes of dissolution within them (e.g. Alsharhan and Kendall, 2003).

Within the net erosion zone, dissolution along groundwater flow paths is the diagnostic characteristic of karst. Most groundwater in the majority of karst systems is of meteoric origin, circulating at comparatively shallow depth and with short residence time underground. Deep circulating, heated waters or waters originating in igneous rocks or subsiding sedimentary basins mix with the meteoric waters in many regions, and dominate the karstic dissolution system in a small proportion of them. At the coast, mixing between seawater and fresh water can be an important agent of accelerated dissolution (Ford and Williams, 2007).

In the erosion zone most dissolution occurs at or near the bedrock surface where it is manifested as surface karst landforms. In a general systems framework most surface karst forms can be assigned to input, throughput or output roles. Input landforms predominate. They discharge water into the underground and their morphology differs distinctly from landforms created by fluvial or glacial processes because of this function. Some distinctive valleys and flat-floored depressions termed poljes convey water across a belt of karst (and sometimes other rocks) at the surface and so serve in a throughput role (Ford and Williams, 2007).

Some karsts are buried by later consolidated rocks and are inert, i.e. they are hydrologically decoupled from the contemporary system. These are referred as palaeokarsts. They have often experienced tectonic subsidence and frequently lie unconformably beneath clastic cover rocks. Contrasting with these are relict karsts, which survive within the contemporary system but are removed from the situation in which they were developed, just as river terraces – representing floodplains of the past – are now remote from the river that formed them. Relict karsts have often been subject to a major change in baselevel. A high-level corrosion surface with residual hills now located far above the modern water table is one example; drowned karst on the coast another. Drained upper level passages in multilevel cave systems are found in perhaps the majority of karsts (Ford and Williams, 2007).

Karst rocks such as gypsum; anhydrite and salt are so soluble that they have comparatively little exposure at the Earth's surface in net erosion zones, in spite of their widespread occurrence. Instead, less soluble or insoluble cover strata such as shales protect them. Despite this protection, circulating waters are able to attack them and selectively remove them over large areas, even where they are buried as deeply as 1000 m. The phenomenon is termed interstratal karstification and may be manifested by collapse or subsidence structures in the overlying rocks or at the surface. Interstratal karstification occurs in carbonate rocks also, but is of less significance. Intrastratal karstification refers to the preferential dissolution of a particular bed or other unit within a sequence of soluble rocks, e.g. a gypsum bed in a dolomite formation (Ford and Williams, 2007).

Cryptokarst refers to karst forms developed beneath a blanket of permeable sediments such as soil, till, periglacial deposits and residual clays. Karst barre´ denotes an isolated karst that is impounded by impermeable rocks. Stripe karst is a barre´ subtype where a narrow band of limestone, etc., crops out in a dominantly clastic sequence, usually with a stratal dip that is very steep or vertical. Recently there has been an emphasis on contact karst, where water flowing from adjoining insoluble terrains creates exceptionally high densities or large sizes of landforms along the geological contact with the soluble strata (Kranjc, 2001).

Karst-like landforms produced by processes other than dissolution or corrosion-induced subsidence and collapse are known as pseudokarst. Caves in glaciers are pseudokarst, because their development in ice involves a change in phase, not dissolution. Thermokarst is a related term applied to topographic depressions resulting from thawing of ground ice. Vulcanokarst comprises tubular caves within lava flows plus mechanical collapses of the roof into them. Piping is the mechanical washout of conduits in gravels, soils, loess, etc., plus associated collapse. On the other hand, dissolution forms such as karren on outcrops of quartzite, granite and basalt are karst features, despite their occurrence on lithologies of that are of low solubility when compared with typical karst rocks (Ford and Williams, 2007).

When there is also a sufficient hydraulic gradient, this can give rise to turbulent flow capable of flushing the detached grains and enlarging conduits by a combination of mechanical erosion and further dissolution. Thus in some quartzite terrains vadose caves develop along the flanks of escarpments or gorges where hydraulic gradients are high. The same process leads to the unclogging of embryonic passages along scarps in sandy or argillaceous limestones. Development of a phreatic zone with significant water storage and permanent water-filled caves is generally precluded. The landforms and drainage characteristics of these siliceous rocks thus can be regarded as a style of fluvio-karst, i.e., a landscape and subterranean hydrology that develops as a consequence of the operation of both dissolution and mechanical erosion by running water (Ford and Williams, 2007).

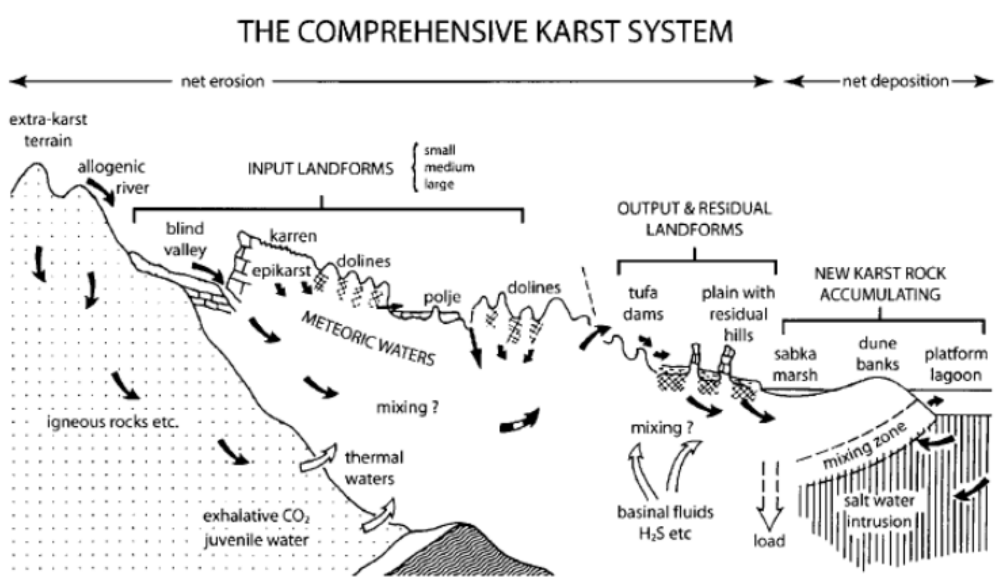


Fig. 1. The comprehensive karst system: a composite diagram illustrating the major phenomena encountered in active karst terrains (Ford and Williams, 2007).

3. Surface features of karst terrains

Since the beginning of karst studies is the surface geology, the surface karst features are the signature of karst performance in the area. Distinguishing and recognition of these phenomenons denote to the development of karst. Different karst features like various types of karrens, dolines (sinkholes), ponors, poljes and springs will introduce and their mechanism of formation will be discussed.

3.1 Karrens

The characteristics of karrens are mainly adopted from Gunn (2004). Limestone that outcrops over large areas as bare and rocky surfaces is furrowed and pitted by characteristic sculpturing landforms that generate a distinctive karstic landscape. These solutional forms, ranging in size from less than 1 mm to more than 30 m, are collectively called karren, an anglicized version of the old German word Karren (the equivalent of the French terms lapiés and lapiaz). Currently, these groups of complex karren forms tend to be called karrenfields or Karrenfelder, in order to differentiate such large-scale exokarst landforms from their smaller karren components (see Table 1).

Several different weathering processes may produce microkarren over limestone surfaces. Some of the microkarren features, such as biokarstic borings, are the result of specific solutional processes induced by cyanobacteria, fungi, algal coatings, and lichens.

At this scale, many different patterns of minute hollows and pits are common, especially in arid environments, because the occasional wetting of the rock produces irregular etching, frequently coupled with biokarstic action. Microrills are the smallest karren form showing a distinctive rilling appearance. Microrills consist of very tiny and sinuous runnels, 0.5–1 mm wide, rarely more than 5 cm long; they are caused by dew and thin water films, enhanced in coastal locations by supralittoral spray. Some other specific karren features develop near the coastline.

The majority of etched surfaces in semiarid environments display a rather complex microtopography that rarely presents linear patterns, the only exception being microrills. The general trend is a chaotic and holey limestone surface in which focused corrosion dominates, without any kind of integration in drainage patterns. These solutional features related to focused corrosion, give rise to depressions of different sizes, more or less circular in plan, such as the rainpit and the kamenitza karren types. Rainpits are small cup-like hollows, sub-circular in plan and nearly parabolic in cross section, whose diameter ranges from 0.5–5 cm and rarely exceed 2 cm in depth; they appear clustered in groups, or even packed by coalescence. The kamenitza karren type (Table 1) consists of solution pans, generally flat-bottomed, from a few square centimeters to several square meters in size, that are produced by the solutional action of still water that accumulates after rainfall; their borders, frequently elliptical or circular in plan, are overhanging and may have small outlet channels.

Many types of karren are linear in form, controlled by the direction of channeled waters flowing along the slope under the effect of gravity. The smaller ones are called rillenkarren and are easy to distinguish from solution runnels or rinnenkarren by their trough width, which rarely exceeds 4 cm. Rillenkarren can be defined as narrow solution flutes, closely packed, less than 2.5 cm in mean width, consisting of straight grooves separated by sharp parallel ribs, that are initiated at the rock edges and disappear downwards. Rillenkarren are produced by direct rainfall and their limited extent seems to be explained by the increase of water depth attaining a critical value that inhibits further rill growth downslope. Neither dendritic patterns nor tributary channels can be recognized in rillenkarren flutes, as opposed to the normal (Hortonian) erosional rills.

Solution runnels are not as straight and regular in form as rillenkarren, being greater and more diversified in shape and origin. Solution runnels or rinnenkarren are normal (Hortonian) rills and develop where threads of runoff water are collected into channels. Classification of solution runnels is difficult because of the great diversity of topographic conditions, the complex processes involved, and the specific kind of water supply feeding the channel. Rinnenkarren is the common term to describe the equivalent of Horton's first-order rills on soluble rocks; they result from the breakdown of surface sheetflows that concentrate into a channelled way and they are also wider than rillenkarren. These solution runnels are sculpted by the water runoff pouring down the flanks of the rocks and have distinctive sharp rims separating the channels; their width and depth range from 5–50 cm, being very variable in length (commonly from 1–10 m, but in some cases exceeding 20 m long). Rundkarren are rounded solution runnels developed under soil cover; they differ from rinnenkarren in the roundness of the rims between troughs and can be considered good indicators of formerly soil-covered karren. Many transitional types from rundkarren to rinnenkarren can be found, due to deforestation and re-shaping of the rocks after subsequent soil removal by erosion. Undercut runnels or hohlkarren are associated with semi-covered conditions, as suggested by the bag-like cross sections of the channel, resulting from enhanced corrosion at the soil contact. Decantation runnels are rills, which reduce in width and depth downslope because the solvent supply is not directly related to rainfall, but corresponds to overspilling stores of water, such as moss clumps, small snow banks, or soil remnants. Wall karren are the typical straight runnel forms developing on sub-vertical slopes, but meandering runnels are more frequent on moderately inclined surfaces or where some kind of decantation feeding occurs over flat areas or gentle slopes. Wall karren may attain remarkable dimensions exceeding 30 m in length. Obviously, transitional forms of runnels are abundant in the majority of karren outcrops, with the exception of areas with arid climates.

Other types of karren features are linear forms controlled by fractures. Grikes or kluftkarren are solutionally widened joints or fissures, whose widths range from 10 cm to 1 m, being deeper than 0.5 m and several meters long. Grikes are one of the commonest and widespread karren features and separate limestone blocks into tabular intervening pieces, called clints in the British literature and Flachkarren in German. For this reason, clint and grike topography is the most typical trend in the limestone pavements, such as the Burren (Ireland; see separate entry) and Ingleborough (northwest England; see Yorkshire Dales entry). The term “cutters” is commonly used in North America as a synonym for grike, although it is best applied to a variety of grike that develops beneath soil cover. Giant grikes, larger than 2 m wide to over 30 m deep, are called bogaz or corridors. Corridor karst or labyrinth karst constitutes the greatest expression of this type of fracture-controlled karrenfield. Splitkarren are similar smaller scale features, resulting from solution of very small weakness planes, being less than 1 cm deep and 10 cm long. Since they conduct water to the karst aquifers, grikes are very important.

Finally, there is a group of karren features closely related to the solutional action of unchannelled washing by water sheets. Many of them, particularly trittkarren and solution ripples, show a characteristic trend that is transverse to the rock slope. At the foot of rillenkarren exposures, subhorizontal belts of unchannelled surfaces can be observed; they are called solution bevels and appear as smoothed areas flattened by sheet water corrosion. More distinctive forms are trittkarren or heelsteps, which are the result of complex solutional processes involving both horizontal and headward corrosion resulting from the thinning of water sheets flowing upon a slope fall. The single trittkarren consists of a flat tread-like surface, 10–40 cm in diameter, and a sharp backslope or riser, 3–30 cm in height.

A wide variety of peculiar karren forms are produced by special conditions, such as where solution takes place in contact with snow patches or damp soil. Trichterkarren are funnel-shaped forms that resemble trittkarren, but are formed at the foot of steep outcrops where

Solutional agent	Karren forms								Synonyms
Biokarstic	Borings								
Weting		Irregular etching							
Tiny water films		Microrills							Rillensteine
Storm shower			Rainpits						Solution pits
Direct rainfall			Rillenkarren						Solution flutes
Chanelled water flow					Solution runnels				Rinnenkarren
						Wall karren			Wandkarren
					Decantation runnels				
					Meandering runnels				Maanderkarren
Standing water				Kamenitza					Solution pans
Sheet wash water flow				Solution bevels					Ausgleichflachen
				Trittkarren					Heelsteps
		Cockling patterns							
			Solution ripples						
Snow melting				Trichterkarren					Funnel karren
				Sharpened edges					Lame dentates
					Decantation runnels				
					Meandering runnels				Maanderkarren
Iced melting						Meandering runnels			
Infiltration					Grikes				Kluftkarren
Soil percolation water					Rundkarren				Rounded runnels
				Smoth surfaces					Bodened runnel, Subcataneous karren
				Subsoil tubes					
				Subsoil hollows					
					Cutters				

Complex processes					Undercut runnels				Hohlkarren
					Clints				Flachkarren
					Pinnacles				Spitzkarren
							Pinnacle karrenfield		Karrenfeld
							Limestone pavement		
								Stone forest	
								Arete karst	
	0-1mm	1mm-1cm	1-10cm	10cm-1m	1-10m	10-100m	100m-1km	>1km	Lapies

Table 1. Classification of karren forms. Light grey areas enclose elementary karst features. Dark grey areas enclose complex large-scale landforms, namely karren assemblages and karrenfield types (Gines et al., 2009)

snow accumulates. Sharpened edges or “lame dentate”, as funnel karren features, are developed beneath snow cover. Rounded smooth surfaces, associated with subsoil tubes and hollows are very common subcutaneous forms, due to the slow solution produced in contact with aggressive water percolating through the soil.

In Bögli’s classifications, two kinds of complex karren forms are recognized: clints or flachkarren, and pinnacles or spitzkarren. These latter, three-dimensional forms, range from 0.5–30 m in height and several meters wide, and are formed by assemblages of single karren rock features, being the constituents of larger-scale groups of complex forms, the karrenfields or karrenfelder. Pinnacles or spitzkarren are pyramidal blocks characterized by sharp edges, resulting from the solutional removal of rock from their sides, as well as from cutting through furrow karren features. Pinnacles are exceptionally well developed in the tropics, where spectacular landscapes constituted by very steep ridges and spikes have been reported. In some cases, such as the Shilin or Stone Forest of Lunan, the presence of transitional forms, evolving from subsoil dissected stone pinnacles sometimes called “dragons’ teeth” to huge and rilled pinnacles more than 30 m in height, can be observed.

Karrenfields are bare, or partly bare, extensions of karren features, from a few hectares to a few hundred square kilometres. Additional work is needed to clarify the relation between karren assemblages and climate, on the basis of the current knowledge accumulated in the last decades from arctic, alpine, humid-temperate, mediterranean, semiarid, and humid-intertropical karsts.

3.2 Sinkhole

Sinkholes are "enclosed hollow of moderate dimensions" originating due to dissolution of underlying bedrock (Monroe, 1970). More specially, sinkholes are surficial landform, found in karst areas and consist of an internally drained topographic depression that is generally circular, or elliptical in plain view, with typically bowel, funnel, or cylindrical shape. Although the circular plan view and funnel shape are ideal forms for sinkholes, they may coalesce into irregular groups or have shapes that are much more complex (Wilson, 1995). The terms sinkholes and dolines are synonymous.

Sinkholes develop by a cluster of inter-related processes, including bedrock dissolution, rock collapse, soil down-washing and soil collapse. Any one or more of these processes can create a sinkhole. The basic classification of sinkholes has six main types that relate to the dominant process behind the development of each, the main characteristics of which are shown in Table 2 and further considered below.

From the lowest point on their rim, their depths are typically in the range of a few meters to tens of meters, although some can be more than a hundred meters deep and occasionally even 500 m. Their sides range from gently sloping to vertical, and their overall form can range from saucer-shaped to conical or even cylindrical. Their lowest point is often near their centre, but can be close to their rim. Dolines are especially common in terrains underlain by carbonate rocks, and are widespread on evaporite rocks. Some are also found in siliceous rocks such as quartzite. Dolines have long been considered a diagnostic landform of karst, but this is only partly true. Where there are dolines there is certainly karst, but karst can also be developed subsurface in the hydrogeological network even when no dolines are found on the surface.

The term sinkhole is sometimes used to refer both to dolines (especially in North America and in the engineering literature) and to depressions where streams sink underground, which in Europe are described by separate terms (including ponor, swallow hole, and stream-sink). Thus the terms doline and sinkhole are not strictly synonymous. Hence, to avoid the ambiguity that sometimes arises in general usage, further qualification is required, such as solution sinkhole or collapse sinkhole. Indeed, the international terminology that is used to refer to dolines that are formed in different ways can also be very confusing. Table 3 lists the terms employed by different authors, the range of terms partly reflecting the extent to which genetic types are subdivided.

The followings are the description of six main types of sinkholes which is described by Waltham and Fookes (2005):

Dissolution sinkholes are formed by slow dissolutional lowering of the limestone outcrop or rockhead, aided by undermining and small-scale collapse. They are normal features of a karst terrain that have evolved over geological timescales, and the larger features are major landforms. An old feature, maybe 1000 m across and 10 m deep, must still have fissured and potentially unstable rock mass somewhere beneath its lowest point. Comparable dissolution features are potholes and shafts, but these are formed at discrete stream sinks and swallow holes, whereas the conical sinkholes are formed largely by disseminated percolation water.

Collapse sinkholes are formed by instant or progressive failure and collapse of the limestone roof over a large cavern or over a group of smaller caves. Intact limestone is strong, and large-scale cavern collapse is rare. Though large collapse sinkholes are not common, small-scale collapse contributes to surface and rockhead degradation in karst, and there is a continuum of morphologies between the collapse and dissolution sinkhole types.

Caprock sinkholes are comparable to collapse sinkholes, except that there is undermining and collapse of an insoluble caprock over a karstic cavity in underlying limestone. They occur only in terrains of palaeokarst or interstratal karst with major caves in a buried limestone, and may therefore be features of an insoluble rock outcrop (Thomas, 1974).

Dropout sinkholes are formed in cohesive soil cover, where percolating rainwater has washed the soil into stable fissures and caves in the underlying limestone (Table 2). Rapid failure of the ground surface occurs when the soil collapses into a void that has been slowly enlarging and stoping upwards while soil was washed into the limestone fissures beneath (Drumm et al, 1990; Tharp, 1999; Karimi and Taheri, 2010). They are also known as cover collapse sinkholes.

Suffusion sinkholes are formed in non-cohesive soil cover, where percolating rainwater has washed the soil into stable fissures and caves in the underlying limestone. Slow subsidence of the ground surface occurs as the soil slumps and settles in its upper layers while it is removed from below by washing into the underlying limestone - the process of suffusion; a sinkhole may take years to evolve in granular sand. They are also known as cover subsidence sinkholes. A continuum of processes and morphologies exists between the dropout and suffusion sinkholes, which form at varying rates in soils ranging from cohesive clays to non-cohesive sands. Both processes may occur sequentially at the same site in changing rainfall and flow conditions, and the dropout process may be regarded as very rapid suffusion. Dropout and suffusion sinkholes are commonly and sensibly described collectively as subsidence sinkholes and form the main sinkhole hazard in civil engineering (Waltham, 1989; Beck and Sinclair, 1986; Newton, 1987). Subsidence sinkholes are also known as cover sinkholes, alluvial sinkholes, ravelling sinkholes or shakeholes.

Buried sinkholes occur where ancient dissolution or collapse sinkholes are filled with soil, debris or sediment due to a change of environment. Surface subsidence may then occur due to compaction of the soil fill, and may be aggravated where some of the soil is washed out at depth (Bezuidenhout and Enslin, 1970; Brink, 1984). Buried sinkholes constitute an extreme form of rockhead relief, and may deprive foundations of stable footings; they may be isolated features or components of a pinnacled rockhead. They include filled sinkholes, soil-filled pipes and small breccia pipes that have no surface expression.

3.3 Polje

Geologically speaking, a polje is a large, karstic, closed depression with a flat bottom often slightly tilted towards the drainage point and surrounded by steep walls and prone to intermittent flooding (Gams, 1978; Prohic et al., 1998). Poljes tend to be areas used for settlement and economic development; they are often the only arable areas in karstic regions where bare rock outcrops predominate with no soil formation. In this sense, polje flooding is poorly understood and requires greater study in order to mitigate its socioeconomic impact. The first step towards taking preventive measures against this phenomenon should be to establish the dynamics and to determine the cause of the flooding, which may be an unusual high supply of surface water and/or groundwater (Lopez-Chicano et al., 2002).

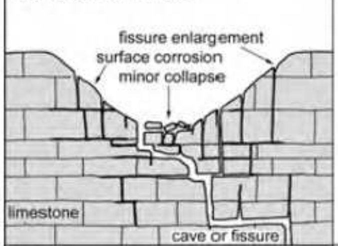
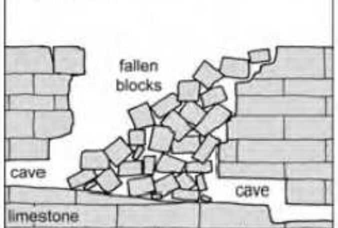
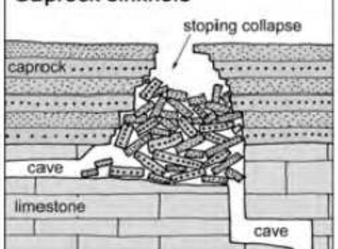
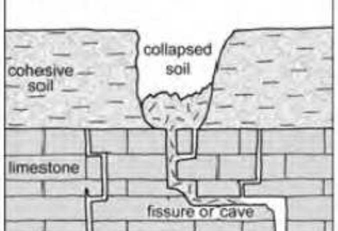
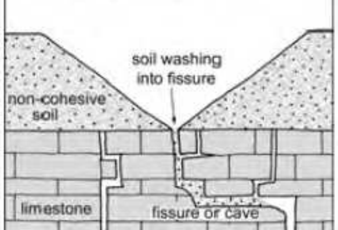
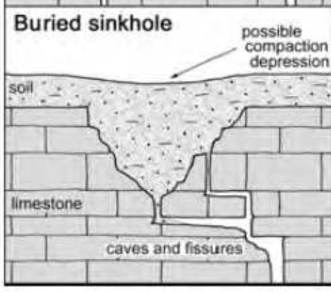
Solution sinkhole	*Formation process*	Dissolutional lowering of surface
	*Host rock types*	Limestone, dolomite, gypsum, salt
	Formation speed	Stable landforms evolving over >20,000 years
	Typical max size	Up to 1,000 m across and 100 m deep
	Engineering hazard	Fissure and cave drains must exist beneath floor
	Other names in use	Dissolution s/h, cockpit, doline
Collapse sinkhole	*Formation process*	Rock roof failure into underlying cave
	*Host rock types*	Limestone, dolomite, gypsum, basalt
	Formation speed	Extremely rare, rapid failure events, into old cave
	Typical max size	Up to 300 m across and 100 m deep
	Engineering hazard	Unstable breakdown floor; failure of loaded cave roof
	Other names in use	Cave collapse s/h, cenote
Caprock sinkhole	*Formation process*	Failure of insoluble rock into cave in soluble rock below
	*Host rock types*	Any rock overlying limestone, dolomite, gypsum
	Formation speed	Rare failure events, evolve over >10,000 years
	Typical max size	Up to 300 m across and 100 m deep
	Engineering hazard	Unastable breakdown floor
	Other names in use	Subjacent collapse s/h, interstratal karst
Dropout sinkhole	*Formation process*	Soil collapse into soil void formed over bedrock fissure
	*Host rock types*	Cohesive soil overlying limestone, dolomite, gypsum
	Formation speed	In minutes, into soil void evolved over months or years
	Typical max size	Up to 50 m across and 10 m deep
	Engineering hazard	The main threat of instant failure in soil-covered karst
	Other names in use	Subsidence s/h, cover collapse s/h, alluvial s/h
Suffosion sinkhole	*Formation process*	Down-washing of soil into fissures in bedrock
	*Host rock types*	Non-cohesive soil over limestone, dolomite, gypsum
	Formation speed	Subsiding over months or years
	Typical max size	Up to 50 m across and 10 m deep
	Engineering hazard	Slow destructive subsidence over years
	Other names in use	Subsidence s/h, cover subsidence s/h, alluvial s/h
Buried sinkhole	*Formation process*	Sinkhole in rock, soil-filled after environmental change
	*Host rock types*	Rockhead depression in limestone, dolomite, gypsum
	Formation speed	Stable features of geology, evolved over >10,000 years
	Typical max size	Up to 300 m across and 100 m deep
	Engineering hazard	Local subsidence on soft fill surrounded by stable rock
	Other names in use	Filled s/h, compaction s/h, paleosinkhole

Table 2. The six types of sinkholes, with typical cross sections and major parameters for each type (Waltham et al., 2005).

Doline-forming processes	Ford and Williams (1989)	White (1988)	Jennings (1985)	Bogli (1980)	Sweeting (1972)	Culshaw and Waltham (1987)	Beck and Sinclair (1986)	Other terms in use
Dissolution	solution	solution	solution	solution	solution	solution	solution	
Collapse	collapse	collapse	collapse	Collapse (fast) or subsidence (slow)	collapse	collapse	collapse	
Caprock collapse			Subjacent collapse		Suffusion subsidence			Interstratal collapse
Dropout	Subsidence	Cover collapse	subsidence	alluvial	alluvial	Subsidence	Cover collapse	
Suffusion	suffusion	Cover subsidence					Cover subsidence	Raveled, shakehole
Burial								Filled, paleosubsidence

Table 3. Doline/sinkhole English language nomenclature as used by various authors (modified from Waltham and Fookes, 2002).

For a depression to be classified as a polje, Gams (1978) identified three criteria that must be met:

1. Flat floor in rock (which can also be terraced) or in unconsolidated sediments such as alluvium;
2. A closed basin with a steeply rising marginal slope at least on one side;
3. Karstic drainage.

He also suggested that the flat floor should be at least 400 m wide, but this is arbitrary because Cvijic´ (1893) took 1km as a lower limit. In fact, poljes vary considerably in size. The floors of reported poljes range from ~1 to > 470 km^2 in area (Lika Polje is the largest at 474 km^2), but even in the Dinaric karst most are less than 50 km^2, and elsewhere in the world a majority are less than 10 km^2 (Ford and Williams, 2007).

Ford and Williams (2007) categorized polje to the three basic types namely border, structural and baselevel poljes.

3.4 Ponor

Concentrated inflows of water from allogenic sources sink underground at swallow holes (also known as swallets, stream-sinks or ponors). They are of two main types: vertical point-inputs from perforated overlying beds and lateral point-inputs from adjacent impervious rocks. The flow may come from: (i) a retreating overlying caprock, (ii) the updip margin of a stratigraphically lower impermeable formation that is tilted, or (iii) an impermeable rock across a fault boundary. A perforated impermeable caprock will funnel water into the karst in much the same way as solution dolines, except that the recharge point is likely to be defined more precisely and the peak inflow larger. Inputs of this kind favour the development of large shafts beneath. Lateral-point inputs are usually much greater in volume, often being derived from large catchments, and are commonly associated with major river caves. The capacity of many ponors in the Dinaric karst exceeds 10m^3/s and the capacity of the largest in Biograd-Nevesinjko polje is more than 100m^3/s (Milanovic, 1993). When the capacity of the swallow hole is exceeded, back-flooding occurs and surface overflow may result (Ford and Williams, 2007).

3.5 Caves

The definition adopted by most dictionaries and by the International Speleological Union is that a cave is a natural underground opening in rock that is large enough for human entry. This definition has merit because investigators can obtain direct information only from such caves, but it is not a genetic definition. Ford and Williams (2007) define a karst cave as an opening enlarged by dissolution to a diameter sufficient for 'breakthrough' kinetic rates to apply if the hydrodynamic setting will permit them. Normally, this means a conduit greater than 5–15mm in diameter or width, the effective minimum aperture required to cross the threshold from laminar to turbulent flow.

Isolated caves are voids that are not and were not connected to any water input or output points by conduits of these minimum dimensions. Such non-integrated caves range from vugs to, possibly, some of the large rooms occasionally encountered in mining and drilling. Protocaves extend from an input or an output point and may connect them, but are not yet enlarged to cave dimensions.

Where a conduit of breakthrough diameter or greater extends continuously between the input points and output points of a karst rock it constitutes an integrated cave system. Most enterable caves are portions of such systems (Ford and Williams, 2007).

Culver and White (2005) present a general cave classification and Palmer (1991) presents a genetic classification of caves.

3.6 Springs

Karst springs are those places where karst groundwater emerges at the surface. Karst spring discharge ranges over seven orders of magnitude, from seeps of a few milliliters per second to large springs with average flows exceeding 20 m^3/s. Flow may be steady, seasonal, periodic, or intermittent, and may even reverse. Karst springs are predominantly found at low topographic positions, such as valley floors, although they may be concealed beneath alluvium, rivers, lakes, or the sea (vrulja). Some karst springs emerge at more elevated positions, usually as a result of geological or geomorphological controls on their position (Gunn, 2004).

Springs in non-karst rocks may result from the convergence of flow in a topographic depression or from the concentration of flow along open fractures such as faults, joints, or bedding planes. Flow in porous media is limited by hydraulic conductivity, so that associated springs almost always have very small flow, often discharging over an extensive "seepage face." Larger springs are possible in fractured rocks such as basalt, where flow may be concentrated along open or weathered fractures. What distinguishes karst springs is that they are the output points from a dendritic network of conduits, and therefore tend to be both larger and more variable in discharge and quality than springs arising from coarse granular or fractured media.

In general, karst springs can be considered in terms of their hydrological function, their geological position, and their karstic drainage or “plumbing”. Karst springs have been classified in many different ways. In theory, different attributes could be combined to describe a spring. For instance the spring at Sof Omar Cave, Ethiopia could be described as a “perennial, full-flow, gravity resurgence”. In practice, most karst springs are described in terms of their most important attribute, depending upon the interest of the observer and the context of the application (Gunn, 2004).

The location and form of karst springs is determined primarily by the distribution of karst rocks, and the pattern of potential flow paths (fractures) in the rock (Karimi et al., 2005). Where karst rocks are intermixed with impermeable rocks, the latter act as barriers to groundwater flow, and karst springs tend to develop as “contact springs” where the boundary between the karst and impermeable rock is exposed at the surface. Where the impermeable unit underlies the karst, it enhances the elevation of the karst water, and the spring (and aquifer) is considered “perched”, as it lies above the topographically optimum discharge point. Where the impermeable unit overlies the karst aquifer, it enhances the pressure of karst water, and springs are then described as “confined perched springs, and so exhibit more sustained flow (Gunn, 2004).

The quality and magnitude of flow from a karst spring reflects the form and function of the karst aquifer, and in particular the recharge processes and the conduit network. Springs deriving much of their water from allogenic surface catchments are known as resurgences. Springs in autogenic aquifers, which receive the bulk of their recharge from a karst surface, are known as exsurgences and they exhibit less variability in discharge and composition. In the past, such flow behaviour has been attributed to distinctive “diffuse”, “conduit” and “pseudo-diffuse” (Karimi et al., 2003) Karst aquifers, but it is now recognized that recharge or underflow-overflow effects are responsible, and that a diffuse karst aquifer is an oxymoron (Gunn, 2004).

A few karst springs show remarkable periodicity in their flow, with a typical period of minutes to hours. In general, this is attributed to the existence of an internal siphon, which progressively fills and drains. Periodicity in hydrothermal springs is seen in geysers. The key feature of geysers is the warming of a pressurized body of water to boiling point and the explosive spontaneous boiling occurring as pressure is released.

Many karst springs occur adjacent to or beneath the surface of rivers, lakes, or the sea; the majority is likely unacknowledged. The interaction between the aquifer and the external water body rests on the hydraulic head distribution and the pattern of connections (springs, sinks, and estavelles) that exists.

Where karst spring water is supersaturated, calcareous tufa deposits develop at the orifice and downstream. Such petrifying springs mantle all objects in calcite, and often build up distinctive mounds and barrages in areas of peak precipitation.

3.6.1 Spring hydrograph analysis

Karst-spring hydrograph analysis is important, first, because the form of the output discharge provides an insight into the characteristics of the aquifer from which it flows and, second, because prediction of spring flow is essential for careful water resources management. However, although the different shapes of outflow hydrographs reflect the variable responses of aquifers to recharge, Jeannin and Sauter (1998) expressed the opinion that inferences about the structure of karst systems and classification of their aquifers is not efficiently accomplished by hydrograph analysis because hydrograph form is too strongly related to the frequency of rainfall events. If a long time-series of such records is represented as a curve showing the cumulative percentage of time occupied by flows of different magnitude, then abrupt changes of slope are sometimes revealed in the curve, which have been interpreted by Iurkiewicz and Mangin (1993), in the case of Romanian springs, as representing water withdrawn from different parts of the karst system under different states of flow. For these reasons, analysis of the recession limbs of spring hydrographs offers considerable potential insight into the nature and operation of karst drainage systems (Bonacci 1993), as well as providing information on the volume of water held in storage. Sauter (1992), Jeannin and Sauter (1998) and Dewandel et al. (2003) provide important recent reviews of karst-spring hydrograph and chemograph analyses (Ford and Williams, 2007).

The principal influence on the shape of the output hydrograph of karst springs is precipitation. Rain of a particular intensity and duration provides a unique template of an input signal of a given strength and pattern that is transmitted in a form modified by the aquifer to the spring. The frequency of rainstorms, their volume and the storage in the system, determines whether or not recharge waves have time to pass completely through the system or start to accumulate. Antecedent conditions of storage strongly influence the proportion of the rainfall input that runs off and the lag between the input event and the output response. The output pattern of spring hydrographs is, however, moderated by the effect of basin characteristics such as size and slope, style of recharge, drainage network density, geological variability, vegetation and soil. As a consequence of all the above, flood hydrograph form and recession characteristics show considerable variety (Ford and Williams, 2007).

Given widespread recharge from a precipitation event over a karst basin, the output spring will show important discharge responses, characterized by:

1. A lag time before response occurs;
2. A rate of rise to peak output (the 'rising limb');
3. A rate of recession as spring discharge returns towards its pre-storm outflow (the 'falling limb');
4. Small perturbations or 'bumps' on either limb although best seen on the recession.

When the hydrograph is at its peak, storage in the karst system is at its maximum, and after a long period of recession storage is at its minimum. The slope of the subsequent recession curve indicates the rate of withdrawal of water from storage. The characterization of the rate of recession and its prediction during drought are necessary for determining storage and reserves of water that might be exploited.

Maillet's exponent implies that there is a linear relation between hydraulic head and flow rate (commonly found in karst at baseflow), and the curve can be represented as a straight line with slope -α if plotted as a semilogarithmic graph. It can be represented in logarithmic form as

$$\log Q_t = \log Q_0 - 0.4343 t\alpha \tag{1}$$

from which α may be evaluated as:0

$$\alpha = \frac{\log Q_1 - \log Q_2}{0.4343(t2 - t1)} \quad (2)$$

Semi-logarithmic plots of karst spring recession data often reveal two or more segments, at least one of which is usually linear (Figure 2). In these cases the data can be described by using separate expressions for the different segments. Jeannin and Sauter (1998) and Dewandel et al. (2003) explain the various models that have been used to try to conceptualize the structure of the karst drainage system that has given rise to the hydrograph form observed and the means by which its recession might be analysed. If the karst system is represented as consisting of several parallel reservoirs all contributing to the discharge of the spring and each with its individual hydraulic characteristics, then the complex recession of two or more linear segments can be expressed by a multiple exponential reservoir model:

$$Q_t = Q_{01}e^{-\alpha 1t} + Q_{02}e^{-\alpha 2t} + \cdots + Q_{0n}e^{-\alpha nt} \quad (3)$$

Milanovic (1976) interpreted the data for the Ombla regime (Figure 2) in Croatia as indicating flow from three types of porosity, represented by the three recession coefficients of successive orders of magnitude. He suggested that α_1 is a reflection of rapid outflow from caves and channels, the large volume of water that filled these conduits emptying in about 7 days. Coefficient α_2 was interpreted as characterizing the outflow of a system of well-integrated karstified fissures, the drainage of which lasts about 13 days; and α_3 was considered to be a response to the drainage of water from pores and narrow fissures including that in rocks, the epikarst and soils above the water table, as well as from sand and clay deposits in caves.

Bonacci (1993) provides a discussion of various causes for changes in the value of recession coefficients.

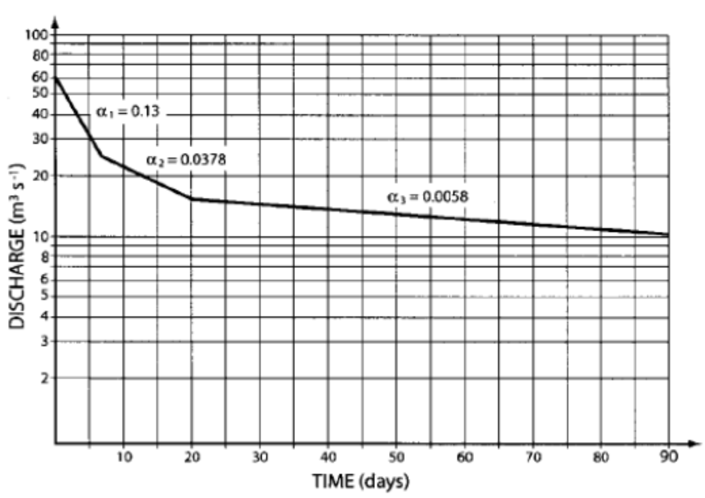


Fig. 2. Composite hydrograph recession of Ombla spring, Croatia (Ford and Williams, 2007).

3.6.2 Quality of karst spring waters

The water emerging from a karst spring consists of a mixture of water from various recharge routes and storage zones. As the environment and duration of recharge, and storage vary, so too will the resulting composition of the water. For example, allogenic recharge water will tend to be more turbid and chemically dilute than autogenic recharge. Long-term storage may result in depletion of the dissolved oxygen in the water, and deep flow may lead to warming or mineralization. In principle, these natural tracers should allow the source and routing of karst spring water to be derived. However, many of these characteristics (e.g. temperature, turbidity, dissolved oxygen, hardness) do not have fixed values associated with particular environments, they are not conserved in transit, and mixing with other waters may induce chemical reactions. The chemical composition of spring waters often results in distinctive deposits, biota, and exploitation, allowing a chemical classification (Gunn, 2004).

3.6.3 Calculation of catchment area and determination of the boundary

The catchment area of each spring is estimated by the following equation:

$$A=Q/(P.I) \tag{4}$$

In which A is the catchment area of the spring (km^2), Q is the total annual volume of water discharging from the spring [million cubic meters (MCM)], P is the total annual precipitation (m) and I is the recharge coefficient (%).

Considering the discharge variations of a spring, one of the following cases is possible:

1. Spring discharge is more or less the same at the beginning and end of the water balance year. In this condition, there are relatively no changes in the system storage and the mean annual discharge of a spring can be used for calculating the annual volume of discharge.
2. If the discharges of a spring at the beginning and end of the water balance year are different; the total discharge of the spring due to the water balance year's precipitations could be estimated by the method proposed by Raeisi (2008).

Determination of recharge coefficient is very difficult. A large number of factors like existence of sinkholes, density of joints and fractures, their openings and type and extent of infillings, percentage, thickness and granulation of soil cover, slope of beds and topography, depth, type, time and space distribution of precipitation, temperature, vegetation cover, etc. can affect the recharge coefficient. Recharge coefficient is defined as the percentage of precipitation, which contributes to the groundwater (spring water). In order to determine the catchment area, it is necessary to have a good approximation of the recharge coefficient for the study area. If there is no idea about the recharge coefficient, the catchment area could be calculated by different recharge coefficients and it will be verified by the proposed method (discussed in the next section) for determining the boundary of a catchment area. Based on experiences in Zagros Mountain Ranges of Iran (Water Resources Investigation and Planning Bureau, 1993; Rahnemaaie, 1994; Raeisi, 1999; Karimi et al., 2001) recharge coefficients can vary between 40 to 90 percent of precipitation. The lower limit is related to areas of low precipitation, high temperature and evaporation and thick soil coverage. The upper limit represents the existence of sinkholes and well-developed karst features.

One of the most complex and difficult problems to deal with in karst hydrology, hydrogeology and geology is the determination of exact catchment boundaries and area of the springs and streamflows (Bonacci and Zivaljevic, 1993). The determination of the catchment boundaries and the catchment area is the starting point in all hydrologic analyses and one of the essential data, which serve as a basis for all hydrologic calculations. In order to exactly define the surface and subsurface catchment boundaries, it is necessary to conduct detailed geologic investigations and, accordingly, extensive hydrogeologic measurements. These measurements primarily involve connections (links) between individual points in the catchment area (connections: ponors-springs, piezometers-piezometers, piezometers-springs) applying one of the tracing methods constituting dye tests, chemical tests, solid floating particles or radioactive matter. The catchment areas in karst vary according to the groundwater levels that change with time. Only in exceptional cases do the surface and subsurface watershed lines coincide and only in those places where the boundaries between catchments are located in impermeable rocks. If this boundary is located in permeable carbonate layers it is not stable.

Figure 3 shows three cases outlining the relationship between a topographic (orographic) and hydrologic (hydrogeologic) spring catchments in the karst. In most cases the basic topographic catchment area A_t is smaller than the hydrologic area A_h, whose boundaries are located within the hydrologic catchment as shown in figure 3A. In practice, it is easy to determine the topographic catchment area, whereas the determination of the hydrologic catchment area is a complex task, difficult to carry out precisely and reliably.

It should be primarily stressed that the definition of the exact catchment area and of the position of the exact catchment boundaries in karst is an interdisciplinary task. It can be exactly and completely carried out only with very close cooperation between various scientists, geologists, hydrogeologists and hydrologists, not excluding the collaboration with the researchers engaged in other scientific disciplines (Karimi, 2003).

It is especially important to define the position of the underground catchment boundaries in karst, and in the analysis of their changes, which is related to the groundwater levels. In order to carry out this task properly it is necessary to install a certain number of piezometers, plan their position and optimum number and to monitor continually the changes of groundwater level in the catchment. In the first phase a small number of piezometers should be installed, and later new ones should be added according to the analysis of the groundwater levels oscillations in the piezometers installed in the first phase.

Estimation of the catchment area in karst can be treated as an inverse problem from the hydrological standpoint. The input vectors are the elements of the water budget (rainfall, inflows, evapotranspiration). Knowing the output vectors (spring discharge or inflow into the river section in karst) it is possible to determine the catchment area, which its position and size ensure the optimum agreement between the hydrograph obtained by calculations and the hydrograph, defined by measurements (Bonacci, 1987, 1990; Bonacci and Zivaljevic, 1993). A promising method in estimation of catchment area is based on the analysis of the groundwater hydrograph, which is caused mainly by groundwater discharge fluctuations. The hydrograph variations, which are dependent on groundwater inflows, describe the outflow from one groundwater reservoir to an open spring. This method can be applied to hydrograph analysis of springs in general and especially to those in karst.

However, according to the basinwide hydrological budget calculations (Degirmenci and Gunay, 1993; Cardillo-Rivera, 2000), hydrological balance and dye tracing tests (Forti et. al, 1990, Karimi et al., 2005), the subsurface catchments were found to be considerably larger than the hydrographic basins

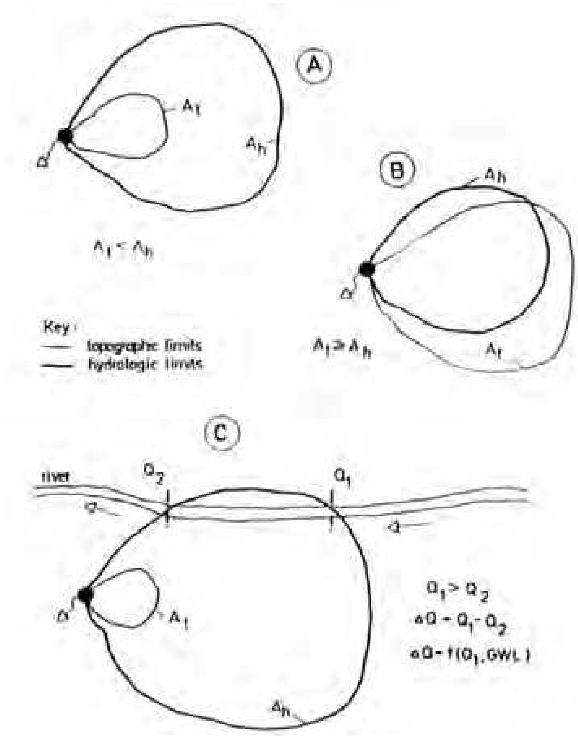


Fig. 3. Three relations between topographic A_t and hydrologic A_h catchment area for karst springs. (A) A_t; (B) A_h; (C) special case when a permanent streamflow is included in the spring catchment along one section (Bonacci, 1987)

Using the calculated approximate catchment area, the most probable location and boundaries of the catchments can be determined by the following procedure:

Step 1. All limestone in the anticline related to the spring and the neighbouring anticlines with higher elevations than the spring is considered as the catchment area.

Step 2. In the area determined in step 1, there must be no hydrogeological and tectonic barriers disconnecting the hydrogeological relationship between the karst aquifer and the spring; In other words, geological and tectonic settings justify the catchment area. Exact and very good geological cross sections could be very useful in this stage. Areas with hydrogeological barriers were disregarded from the catchment area.

Step 3. A general water balance is considered for the area determined in the step 2; i.e. all the outputs (including discharge to alluvium) and also all the inputs will be taken into account, so that the catchment area of the spring does not interfere with the other springs.

The catchment area is probably as close as possible to the spring, i.e. at first, the catchment area is supplied by the spring related anticline.

Step 4. The following parameters are useful in confirming of the determined area:

1. The physico-chemical parameters of the spring display the characteristics of the related karst aquifer and adjacent formations.

2. Isotope studies can be used quantitatively or qualitatively in the determination of the elevation ranges that have a main role in the recharging of the spring.
3. The general direction of flow may be determined using water table elevations or isopotential maps, if piezometers are constructed in the study area.
4. Flow coefficient (FC) parameters is a good representations of the springs which are recharging from the outside the surface catchment area (FC≥ 1). It is defined as the ratio of volume of discharge to the volume of precipitation on the surface catchment area of a given hydrometric station.
5. The normalized base flow (NBF) diagram, which is a useful tool, is used for checking the calculated catchment areas (Connair and Murray, 2002; White, 2002).
6. The characteristics of the catchments such as soil cover, vegetation, sinkholes, morphology, amount and type of precipitation and elevation could be applied to differentiate between different catchment areas.

Step 5. Tracing and geophysics: The catchment areas determined with a high uncertainty about them, could be verified using tracing and geophysical tests. Because of the uncertainty in recharge coefficient, the error in the determined catchment area could be as high as 10 percent.

The catchment area of the Alvand Basin springs, west of Iran, was calculated based on equation 4, and according to the above criteria the most probable boundary of the catchment area of the main karst springs was determined (Figure 4).

3.6.4 Time variations of physico-chemical parameters

Temporal variations of physico-chemical parameters of karst springs have been used to determine aquifer characteristics in the last four decades. Jakucs (1959) showed that the chemical composition and discharge of karst spring water might vary with time. Garrels and Christ (1965) categorized the flow in karst regions into open and closed systems, based on the amounts of CO_2 available for dissolution. White and Schmidt (1966) and White (1969) classified the flow in karst aquifers into conduit and diffuse flow. Karimi et al. (2003) introduced pseudo diffuse flow regime. Shuster and White (1971, 1972) concluded that the type of flow (diffuse or conduit) can be determined by its chemograph. Ternan (1972), Jacobson and Langmuir (1974), Ede (1972), Cowell and Ford (1983) evaluated the flow systems in karst formations using other criteria, such as type of recharge, electrical conductivity and coefficient of variation of electrical conductivity, coefficient of variation of discharge and temperature, and time variations of these parameters.

Based on the above ideas, in a diffuse flow system, recharge is generally conducted through a network of numerous small joints and fractures that are distributed in the karst aquifer. The openings of these fractures are smaller than one centimetre and water slowly reaches the groundwater in a laminar manner. One of the main peculiarities of these aquifers is the small variation of physical and chemical properties of the discharging springs. Natural discharge from such a system is usually through a large number of smaller springs and seeps. In a conduit flow system, the aquifer is fed through either large open fractures (ranging from one centimetre to more than one meter) or sinkholes. In such systems, water reaches the groundwater very quickly and ultimately the springs in a turbulent manner. Hence, the physico-chemical properties of the spring waters are non-uniform. In this type of system, the discharge usually occurs through one single large spring.

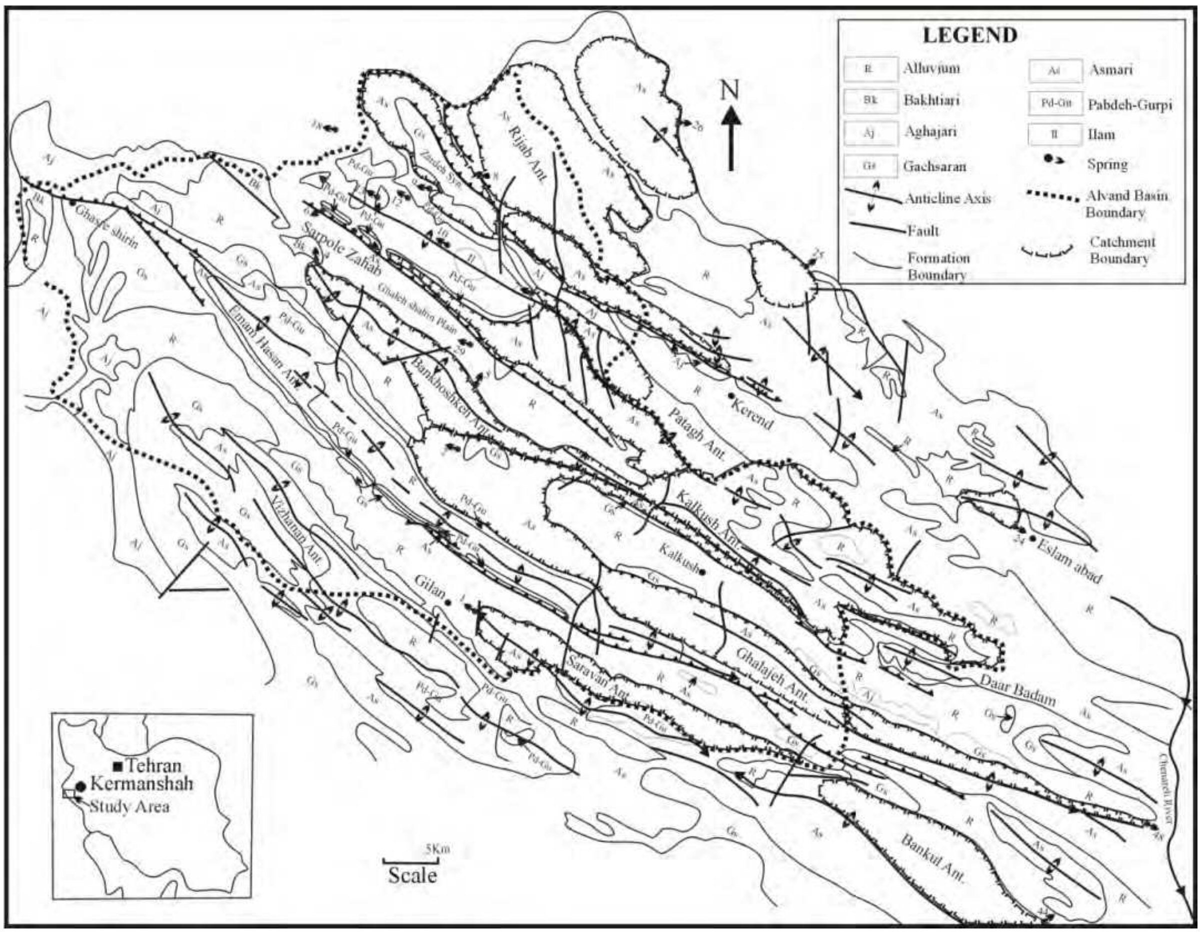


Fig. 4. Boundary of catchment area of the Alvand main karst springs (Karimi et al., 2005).

Bakalowicz (1977), Atkinson (1977), Scanlon and Thrailkill (1987), and Raeisi et al. (1993) were not able to use the criteria proposed by previous workers to determine the flow regime and found contradictory results. The reason is probably due to the fact, that purely diffuse or purely conduit flow systems rarely occur in nature, rather it is a combination of these two types of flow that usually prevail. Raeisi and Karami (1996) suggested that when the physico-chemical characteristics of a karst spring are to be used to determine the properties of the related aquifer, the first step should be the evaluation of the effects of external factors on the outflow. Lopez-Chicano et al. (2001) analyzed the hydrogeo-chemical processes in waters of Betic Cordilleras in Spain by studying hydrography, temporal evolution of physico-chemical parameters, ionic ratios (mainly Mg/Ca) and by means of simple and multivariate statistical analysis. They concluded that the aquifer exhibits diffuse flow.

Time series variations of physico-chemical parameters of springs were inspected by different researchers like Ashton (1966); Hess and White (1988, 1993); Bakalowicz et al. (1974); White (1988, 2002); Williams (1983); Scanlon and Thrailkill (1987); Scanlon (1989); Sauter (1992); Ryan and Meiman (1996); Raeisi and Karami (1996, 1997); Lopez-Chicano et al. (2001) and Desmarais and Rojstaczer (2002). Generally in a typical karst system, after an intense rainfall, discharge increases within a short period and then decreases slowly. In this period, the EC shows an increasing-decreasing-increasing trend. Based on electrical conductivity response, Desmarais and Rojstaczer (2002) divided spring response into three stages. The three stages include flushing, dilution, and recovery.

Flushing: The flushing stage marks the initial response in the spring to storms. The beginning of this stage is signalled by the increase of the slope of the conductivity curve of the spring. There are two hypotheses concerning the water source that causes this flushing of the spring:

1. The flushed water is water that has interacted or equilibrated within the soil zone, and possibly resides in small pores or fractures near the land surface, i.e. the subcutaneous zone. This water would be relatively warm and would likely contain dissolved salts (or would dissolve salts from the soil during transit), which would give the water a relatively high conductivity. The warmer, high electrical conductivity water would be mobilized by the rainwater infiltrating into the soil and pushed toward the spring.
2. The new rainwater is able to rapidly recharge the aquifer, possibly through fractures or surface swallets, and it mobilizes older, deeper water that has been residing in smaller fractures and pores out of the aquifer. This 'old water' is at or near equilibrium with the limestone, but the new water is not. The old water, because it has resided in the aquifer for a relatively long time, would have higher electrical conductivity than the baseflow spring water. Flushing is not typical of all carbonate springs (Ryan and Meiman, 1996; Desmarais and Rojstaczer, 2002).

Dilution: The dilution phase begins with the peak in the electrical conductivity (EC) curve and ends when the EC reaches its minimum value. The start of the conductivity decrease represents the first arrival of storm water at the spring. During this phase, the temperature commonly levels off and then remains constant until the next storm whereas discharge continues to decrease. The area of the recharge basin is the main factor controlling the length of this phase. After that time period, the spring begins to 'recover' because there is very little recharge water remaining.

However, the system response can also be explained by a competition between the velocity at which recharge water is moving through the system, how fast this 'new' water dissolves carbonates to gain the same chemistry signature as the 'old' aquifer water, and the amount of mixing that takes place between these two water sources.

Recovery: The recovery phase begins when the minimum is obtained in conductivity. During this phase, conductivity increases steadily until the next storm begins. The concentrations of all the major cations and anions increase during this period. All of these changes indicate that the system is returning to equilibrium conditions. The conductivity minimum likely indicates that the last of the recharge water has been in contact with the aquifer rock long enough to begin to dissolve limestone and/or dolomite in sufficient quantities to allow the overall system to begin to recover from the dilution. Figure 5 shows the three above-mentioned stages in the SS-5 spring in Bear Creek Valley, Tennessee (Desmarais and Rojstaczer, 2002).

The minimum of electrical conductivity corresponds to the maximum dilution of groundwater by fresh recharged water, which could be used as a representative of lag time of the system. The lag time is a measure of the length of time required for the arrival of unsaturated water (minimum of EC) to reach to the recording station (Hess and White, 1988).

Hess and White (1993) stated that fluctuation in hardness fit the well-established concept that hardness variability is an indication of conduit karstic drainage system as has been observed for many conduit karst aquifers in North America and Europe (Pitty, 1966; Ternan, 1972; Atkinson, 1977).

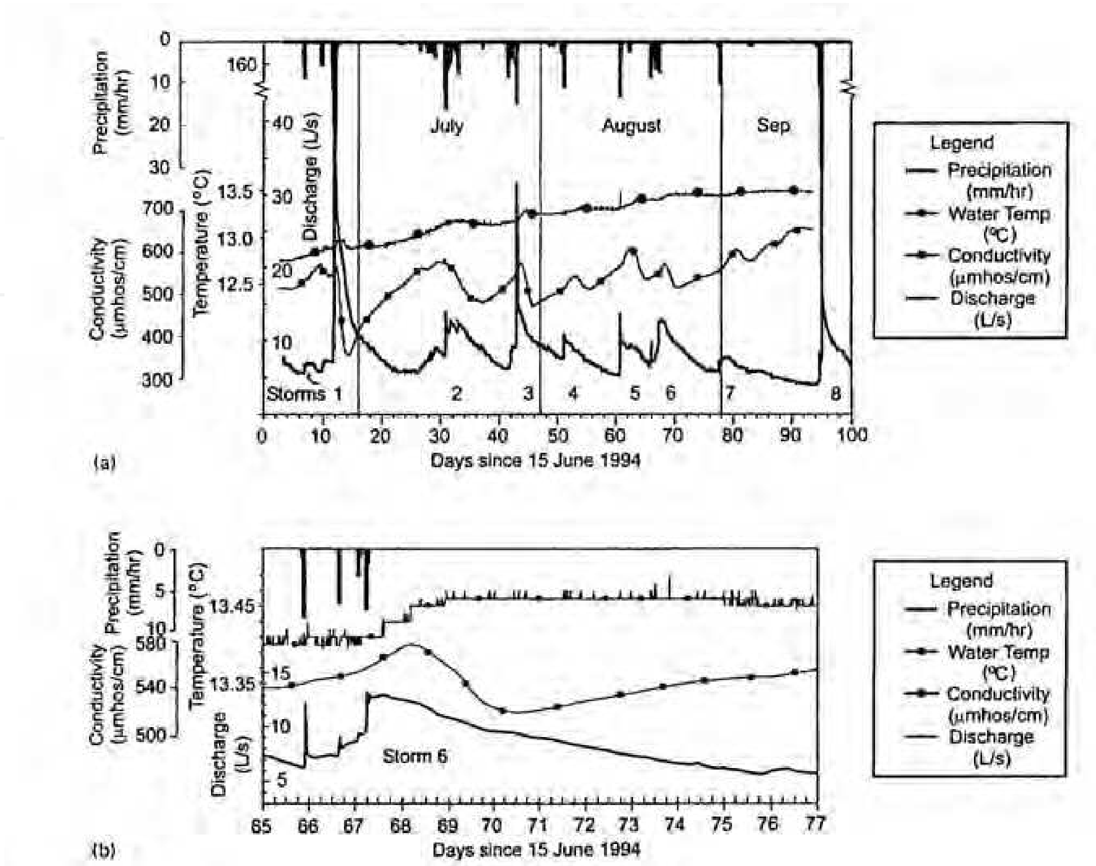


Fig. 5. (a) Discharge, conductivity and temperature at SS-5 for storms 1-8. (b) Detailed Discharge, conductivity and temperature at SS-5 for storm 6 (Desmarais and Rojstaczer, 2002).

Conductivity measurements of the spring water provide an inexpensive and rapid method of distinguishing the rock type through which groundwater flows. The amount of variability of the data through time also gives insight to how rapidly the water quality changes by recharge events. Springs with high conductivity and low coefficient of variation of the data suggest slow groundwater movement through a non-karst aquifer. The data with great variability indicates that groundwater flow is rapid through conduits (Ogden et al., 1993).

3.6.5 Environmental isotopes in hydrogeology

Environmental isotopes now routinely contribute to hydrogeological investigations, and complement geochemistry and physical hydrogeology. Meteoric processes modify the stable isotopic composition of water, and so the recharge waters in a particular environment will have a characteristic isotopic signature. This signature then serves as a natural tracer for the provenance of groundwater. On the other hand, radioisotopes decay, providing us with a measure of circulation time, and thus groundwater renewability. Environmental isotopes provide, however, much more than indications of groundwater provenance and age. Looking at isotopes in water, solutes and solids tells us about groundwater quality, geochemical evolution, recharge processes, rock-water interaction, the origin of salinity and contamination processes (Clark and Fritz, 1997).

Given the relationship between δ18O in precipitation and elevation, it is possible to determine an approximate mean recharge elevation for springs. An inherent assumption in this type of analysis is that the spring water is young enough to be comparable to modern precipitation (James et al., 2000). On the basis of the relationship established between the δ18O value in rainwater and altitude, Kattan (1997); Abbott et al. (2000); James et al. (2000); Yoshimura et al. (2001); Eisenlohr et al. (1997); Vallejos et al. (1997) and Ellins (1992) estimated the mean elevation of recharge zones of groundwater in their study areas. These estimations corresponded more or less to the natural topographic divide line in their study areas.

Even at the same altitudes, there are variations in the δ18O values of respective rains because stable isotope compositions of water vapour masses are different from one to another. Consequently, the mean values of δ18O of rainwater for about a half or full year are plotted against the altitude (Abbott et al., 2000; Yoshimura et al., 2001). Altitude isotope effects were observed in different parts of the world. For example Kattan (1997) reported a figure of -0.23 ‰ per 100 meter increase in elevation in Syria and Abbott et al., (2000) a figure of -0.25 ‰ per 100 meter in USA (Vermont) and other researchers like James et al., (2000); Yoshimura et al. (2001); Vallejos et al. (1997) and Williams and Rodoni (1997) figures of -0.18‰, -0.15 to -0.25‰, -0.35‰ and -0.1‰ per 100 m increase in elevation in USA (Oregon), a tropical area, Spain and California respectively. Leontiadis et al. (1996) provided an estimate of -0.44‰ 100 m^{-1} for the groundwater altitude effect on the δ18O value in Eastern Macedonia and -0.21‰ in Thrace in Greece. Figure 6 shows the determination of recharge elevations from the relationship between snow δ18O and elevation. Approximate recharge elevations can be determined by extrapolating from the spring isotopic composition to the regression line, then dropping a perpendicular line to intersect the abscissa. Inset is the precipitation data used to constrain the regression line (James et al., 2000).

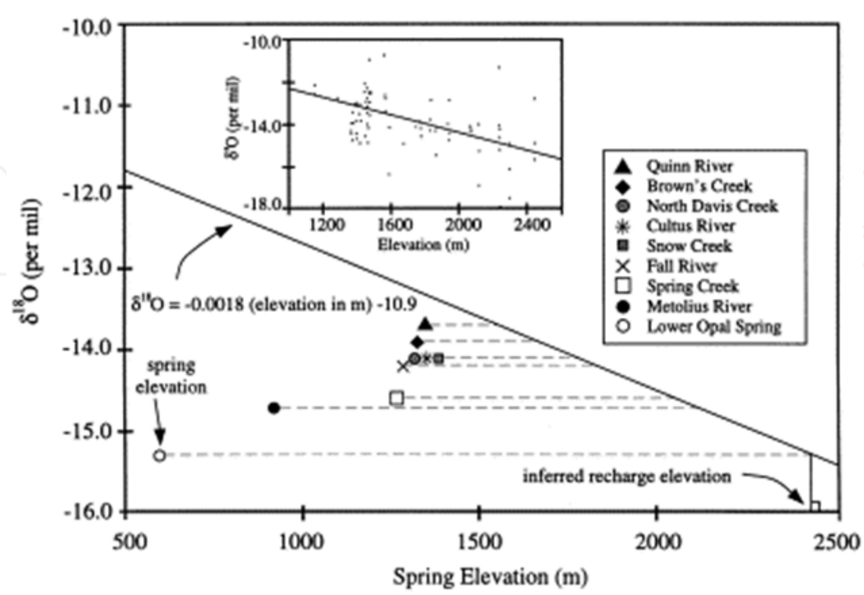


Fig. 6. Determination of recharge elevations from the relationship between δ18O in snow and elevation (James et al., 2000).

Stewart and Williams (1981) stated that the large seasonal variations in $\delta^{18}O$ values in the recharge zones diminish to within experimental error at the springs due to diffusion in the very large groundwater reservoir, thus little information on the rate of groundwater flow can be obtained from these data. Ferdrickson and Criss (1999) found that the isotope variations for the Meramec River Basin (Missouri) springs share the overall, cycloid-like shape of the precipitation, with higher $\delta^{18}O$ values during the summer and fall and excursions to more negative values during the winter months. The $\delta^{18}O$ variations for the precipitation have amplitude exceeding 10‰, yet the annual amplitude of the variations in the rivers was only 3‰ and the amplitudes of several karst springs were even smaller, valuing at about 1‰ (Figures 1-10 and 1-11). They used these seasonal variations to determine the residence time of water.

Isotope studies indicate that there is some dilution in spring water during the rain season because of the mixing of event water with the spring and the enrichment in the dry season (Stewart and Williams, 1981; Ford and Williams, 1989; Ferdrickson and Criss, 1999; Abbot and Bierman, 2000).

4. Different zones of a karst aquifer

Generally an unconfined aquifer can be subdivided into two saturated and unsaturated zones. It is not necessary that all parts of this classification be present in any given karst (Karimi, 2003).

In the unsaturated zone above the water table, voids in the rock are only partially occupied by water, except after heavy rainfall when some voids fill up completely. Water is percolated downward in this zone by a multiphase process, air and water co-existing in the pores and fissures. Because of the high concentration of CO_2 in the soil cover, the percolating water in the soil cover is usually unsaturated and aggressive. When water leaves the soil cover, it behaves as in a closed system because of the lack of any connection to the atmosphere (Ford and Williams, 1989). The subcutaneous zone is situated under the soil cover or in the upper part of the limestone aquifer.

The water-saturated zone below the water table is the phreatic zone (White, 1988). If the karst mass is large and deep enough, it is possible to distinguish a shallow and a deep phreatic zone. The circulation of groundwater is fast in the former and slow in the latter. Different scientists estimated the depth of the shallow phreatic zone to be between 20 to 60 m (Bogli, 1980).

The lower boundary of an aquifer is commonly an underlying impervious formation. But should the karst rocks be very thick, the effective lower limit of the aquifer occurs where no significant porosity has developed. Figure 7 shows alternative approaches to estimating the thickness of an unconfined karst aquifer.

5. Problems of construction in karst regions

Karst processes and landforms pose many different problems for construction and other economic development. Every nation with karst rocks has its share of embarrassing failures such as collapse of buildings or construction of reservoirs that never held water. Prevention of unanticipated remedial measures in karst terrains now imposes a lot of economic problems each year on the governments.

5.1 Rock slide-avalanche hazards in karst

A landslide or rock slide-avalanche is the catastrophically rapid fall or slide of large masses of fragmented bedrock such as limestone (Cruden, 1985). 'Landslide' is more widely used but is also applied to slides of unconsolidated rocks. Rock slides take place at penetrative discontinuities, the mechanical engineering term for any kind of surface of failure within a mass. Once initiated, there is powerful momentum transfer within the falling mass and it may partially ride on a cushion of compressed air that can permit it to run for some hundreds of metres upslope on the other side of a valley (van Gassen and Cruden, 1989).

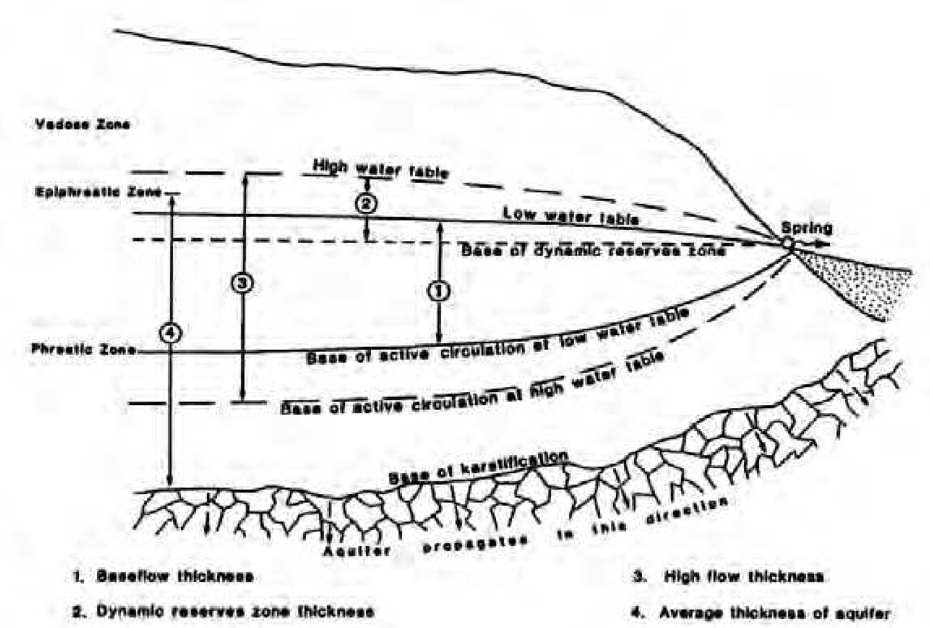


Fig. 7. Alternative approaches to estimating the thickness of an unconfined karst aquifer (Ford and Williams, 1989).

Carbonate rocks and gypsum are especially failure-prone for two reasons.

1. While faults and joints are the only important penetrative discontinuities in most other rocks, in karst strata there is also major penetration via bedding planes. In fact they are particularly favoured as surfaces of failure because of their great extent.
2. Large quantities of water may pass rapidly through the rock via its karst cavities to saturate or lubricate interlaminated or underlying weak or impermeable strata such as clays. The forces that resist catastrophic failure within a particular rock are defined by an internal angle of friction. Minimum angles for relatively hard carbonates without shale interbeds range from 14° to 32°.

The principal settings of landslides in karst rocks are shown in Figure 8. Slab slides are particularly common because they are bedding-plane failures. They are especially frequent and dangerous in the overdip situation. The biggest one in the world is the Saymareh landslide in western Iran (Karimi, 2010). Rotational failures within massive carbonates are comparatively rare but there are large ones in dolomites in the Mackenzie Mountains, Canada. Toppling cliffs are common in all rocks; see Cruden (1989) for formal analysis. Toppling or rotational failures are quite common along escarpment fronts where the permeable karst rock rests on a weak but impermeable base such as a shale; Ali (2005) describes 12 limestone failures of up to 800 × 10^6 t each along a 20 km frontage near the city of Sulaimaniya in northern Iraq, caused by spring sapping at the contact with underlying shales.

Downslope detachment and creep of karst rock formations resting on slick but impermeable strata beneath them (Figure 8) may proceed slowly for long periods and then suddenly accelerate into a landslide, usually as a consequence of heavy rains or an earthquake. At the Vajont Dam disaster of 1963 in the Italian Alps, in which 2000 lives were lost; rise of water level in a reservoir may have contributed by increasing pore water pressures on the slide plane. The Ok Ma Landslide (Papua New Guinea) was a slide of ~ 36×10^6 m^3 of fractured massive limestone on clay dipping into a river valley that was induced by removal of the toe of a previous slide in order to install a dam for a gold mine.

5.2 Setting foundations for buildings, bridges, etc.

Setting foundations where there are soils, etc., covering maturely dissected epikarst can encounter many problems. Figure 9 illustrates the range of different methods that are used to overcome them by compacting the soil or pinning the footings to (comparatively) firm bedrock. Under large or heavy structures the majority of these methods can be very expensive. Reinforced concrete slabs ('rafts' floating on the soil following its mechanical compaction) are now much used as alternatives under buildings. For roads on mantled epikarst or spanning infilled solution and suffusion dolines strong synthetic plastic sheeting, strips or meshes ('geofabrics') are being substituted because they are cheaper: their longterm reliability is not yet established, however. Much has been written on these subjects; see Beck (2005) and Waltham et al. (2005) for recent surveys.

Building calamities remain frequent worldwide.

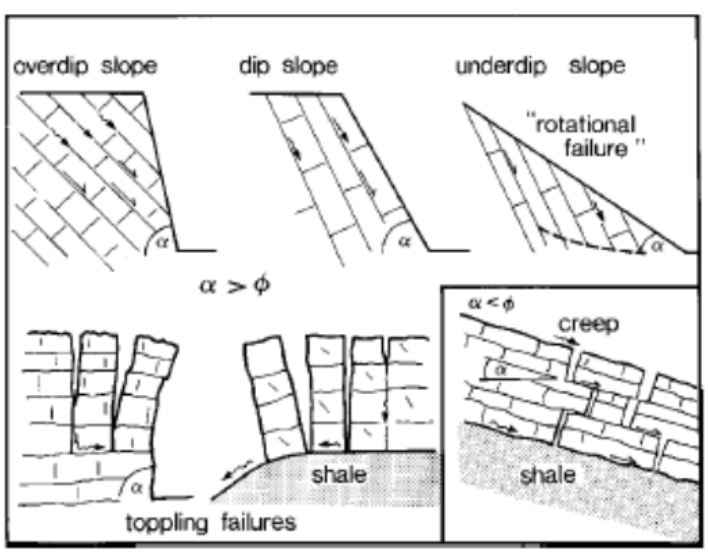


Fig. 8. Types of landslides (or rock slide-avalanches) in carbonate rocks. Ø is the internal angle of friction of the rock. Failures on dip and overdip slopes are termed 'slab slides' (Ford and Williams, 2007).

Cavities entirely within bedrock can also pose dangers if they are at very shallow depth or if the planned structural load is considerable. For typical strong limestones with caves, Waltham et al. (2005) recommend a minimum of 3m bedrock above a cavity 5m wide, 7m for widths >10m; for chalk and gypsum, at least 5m of rock above a cave 5m wide.

Construction on gypsum requires particular care. Gutierrez (1996) and Gutierrez and Cooper (2002) discuss the rich example of Calatayud, a town of 17000 persons in the Ebro Valley, Spain. It is built on a fan of gypsiferous silts interfingering with floodplain alluvium, and underlain by a main gypsum formation ~500m thick. The existing buildings are 12th century to modern in age. Many (of all ages including recent) display subsidence damage that ranges from minor to very severe. The primary cause is believed to be dissolution of the gypsum bedrock, which is abetted by local compaction of overburden accumulated since the town was founded in AD 716, collapse of some abandoned cellars and dissolution of the silts.

5.3 Tunnels and mines in karst rocks

Tunnels and mine galleries (adits or levels) will be cut through rocks in one of three hydrogeological conditions:

(i) vadose; (ii) phreatic but at shallow depth or where discharge is limited, so that the tunnel serves as a transient drain that permanently draws down the water table along its course; (iii) phreatic, as a steady-state drain, i.e. permanently water-filled unless steps are taken drain it. Long tunnels in mountainous country may start in the vadose zone at each end but pass into a transient zone, or even a steady-state phreatic zone, in their central parts.

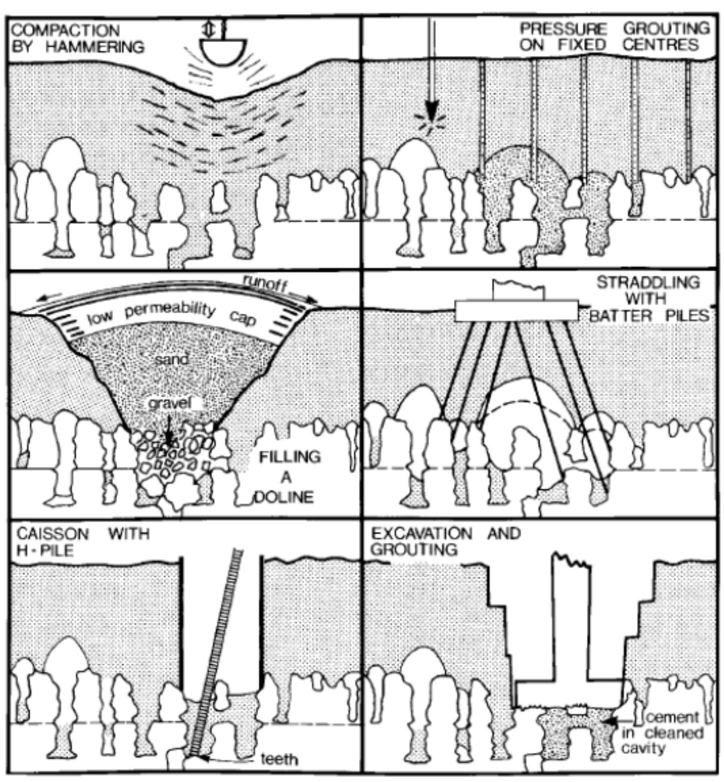


Fig. 9. Illustrations of some of the principal types of foundation treatments in a soil-mantled karst (Ford and Williams, 2007).

Vadose and transient zone tunnels are cut on gentle inclines to permit them to drain gravitationally.

Where the tunnel or mine is a deep transient drain or is in the steady-state phreatic zone, gravitational drainage will not suffice, e.g. if the tunnel is below sea level. Three alternative strategies can then be adopted. The first is to pump from the tunnel itself, when necessary. It is prone to failure if the pumps fail and to disaster (for the miners) if large water-filled cavities are intercepted, causing catastrophic inrushes of water. The second means is to grout the tunnel and then to pump any residual leakage as necessary. It is the essential method for transportation tunnels. Traditionally, tunnel surfaces were rendered impermeable by applying a sealant (e.g. concrete) as they became exposed. This does not deal with the catastrophic inrush problem.

Modern practice is to drill a 360° array of grouting holes forward horizontally, then blast out and seal a section of tunnel inside this completed grout curtain. This largely deals with the hazard of catastrophic inrush, i.e. a flooded cavity should be first encountered by a narrow bore drill hole that can be sealed off quickly. Milanovic (2000) and Marinos (2005) discuss tunnel protection thoroughly, with many examples.

Grouting is not feasible in the extracting galleries of a mine. Here, a third and most elaborate strategy is to dewater the mine zone entirely, i.e. maintain a cone of depression about it for as long as the mine is worked.

A largely debated issue is related to the engineering aspects of karst. As population density increases, need for construction of roads, various infrastructures, and water resources increases. This leads to reclamation projects with construction of dams and reservoirs in or nearby karst regions. The understanding and evaluation of environmental impacts of such human activities on karst are important to try and find a balance between development and preservation of these complex hydrogeological systems (Milanovic, 2002, 2004).

While the problems associated with the construction of a dam site on a karst area are fairly well understood (Uromeihy, 2000; Romanov et al., 2003, 2007; Turkmen, 2003; Xu and Yan, 2004; Ghobadi et al., 2005) consequences of flooding of karst discharge areas due to reservoirs built close to karst regions are much less studied (De Waele, 2008).

5.4 Dam construction on carbonate rocks

A large number of dams have been built on karstic limestones and dolomites for different purposes all over the world. Flood control has been particularly important on branches of the Mississippi, where the Tennessee Valley Authority (TVA) was very active in the first half of the 20th Century. Storage to sustain paddy fields, counter prolonged dry seasons or general drought was more important in China, around the Mediterranean and in Iraq, Iran and other semi-arid areas. Hydroelectric power generation was an early priority in alpine sites and is now the principal goal of perhaps the majority of the larger, higher dams. Few nations that constructed them escaped serious problems due to karst leakage, leading to considerable overruns in cost or to outright abandonment in some instances. These are summarized in many engineering design and construction reports. The TVA main report (1949) is still pertinent; Soderberg (1979) gave a more recent review of their work. Therond (1972), Mijatovic (1981), Nicod (1997) and Milanovic´ (2000, 2004) have discussed European experience, which generally has been with geologically more complex mountainous sites.

Therond (1972) identified seven different major factors that may contribute to the general problem. These are: type of lithology, type of geological structure, extent of fracturing, nature and extent of karstification, physiography, hydrogeological situation and the type of dam to be built. For each factor, clearly, there are a number of significantly different conditions. In Therond's estimation, these together yield a combination of 7680 distinct situations that could arise at dam sites on carbonate rocks! It follows that dam design, exploration and construction must be specific to the particular site, and be continually re-evaluated:

Milanovic´ (2000) suggests that there have been three principal settings for dams on carbonate rocks:

1. In the narrow gorges typically created where large allogenic rivers cross them in steep channels. Here rates of river entrenchment have usually been faster than karst development; with the result that karstification is not a major problem beneath the channels. It may however be hazardous in the gorge walls, which will form the dam abutments.
2. Dams and reservoirs in broader valleys where the karst evolution has been as fast as or faster than river entrenchment. The TVA sites are examples. This can cause many problems beneath the dam as well as in the abutments and upstream in valley sides and bottoms. It is particularly hazardous where the valley is hanging at its mouth, as is common in alpine topography, because the natural (pre-dam) groundwater gradient is steepened there. Unfortunately, this will also be an optimum site for hydroelectric power dam location because the reservoir volume, fall height and gradient of the penstock are all maximized there.
3. In poljes to control flooding and store water for dry season irrigation. This is perhaps the most difficult setting because under natural conditions the dry season water table will be deep below the polje floor in highly karsted rock. The reservoir floor must be sealed (with clay, shotcrete, PVC, etc.) to retain water but the seals can be blown by air pressure as floodwaters rise in the caves underneath. Ponors must be plugged or walled off by individual dams rising above the reservoir water surface, and estavelles must be fitted with one-way valve systems. Much success has been achieved in the poljes of former Yugoslavia but the Cernic´a project there and Taka Polje in Greece are two examples that were abandoned after expensive study.

Many dams are more than 100 m in height and some exceed 200 m. A first, obvious danger of dam construction is that by raising the water table to such extents, an unnaturally steep hydraulic gradient is created with unnatural rapidity across the foundation and abutment rock, and an unnaturally large supply of water is then provided that may follow this gradient. This is a hazardous undertaking because, unless grout curtains penetrate well into unkarstified rock, the increased pressure will drive groundwater movement under the dam and stimulate dissolution. Dreybrodt et al. (2002, 2005) have approached this problem with realistic modeling scenarios for limestone and gypsum. In the limestone case solution conduits are shown to propagate to breakthrough dimensions (turbulent flow) beneath a 100m deep grout curtain under a dam in approximately 80 years. Remedial work would then be essential. Table 4 provides details of leakage from dams in karst before and after remedial works, and Figure 10 shows one example of increasing leakage over time at a dam in Macedonia.

While leakage through dam foundations and abutments is most feared, it is quite possible that there may be lateral leakage elsewhere in a reservoir. Problems with karst can arise even where the dam itself is built on some other rock, if karst rocks are inundated upstream of it. Montjaques Dam, Spain, was built to inundate a polje. It failed by leaking through tributary passages and the scheme was abandoned (Therond, 1972).

In tackling dams on karst the first essential is drilling of exploratory boreholes (with rock core extracted for inspection), and mining of adits (galleries big enough for human entry and inspection) in the abutments. These may later be used for grouting. Surface, downhole and interhole geophysics (Milanovic´, 2000) can amplify the picture but are not in themselves sufficient because they will rarely detect smaller cavities, or even large ones below ~50m or so. Even intensive drilling and mining may be inadequate. At the Keban Dam site in Turkey, despite 36 000m of exploratory drilling and 11 km of exploratory adits, a huge cavern of over 600 000m^3 was not detected; 'expect the unexpected!' (Milanovic´, 2000).

Grout curtains are essentially dams built within the rock. 'Due to karst's hydrogeological nature, grout curtains executed in karstified rock mass are more complex and much larger than curtains in other geological formations' (Milanovic 2004). The surest principle is to grout entirely through the limestone into underlying impermeable and insoluble strata where this is possible. Curtains in abutments can also be terminated laterally against such strata (the 'bathtub' solution).

The normal practice is to excavate all epikarst and fill any large caverns discovered by the adits and bores, then place a main curtain beneath the dam, in the abutments and on the flanks. A cut-off trench and second, denser curtain may be placed upstream in the foundation if there are grave problems there. In the main curtain a first line of airtrack grout holes will be placed on centers never more than 8–10m apart and filled until there is back pressure. A second, offset line of holes is then placed and filled between them. Third and fourth lines may be used until the spacing reaches a desirable minimum that is normally not more than 2m. Adits in the abutments that are used to inject grout should be no more 50m apart vertically. Standard grouts are cement with clay (particularly bentonite, a clay that expands when wetted), plus sand and gravel for large cavities. Mixtures are made up as slurries with differing proportions of water. Ideally, the goal of all grouting is to reduce leakage of water to one Lugeon unit (Lu=1 L $min^{-1}m^{-1}$ of hole at 10 bars water pressure) under a dam and 2 Lu in the abutments. In practice, in karsted limestones it is often difficult to inject grout where permeability is <5.0 Lu. Correlation between Lugeon measured during exploration and the amount of grout that will be required can be very poor also; at Grancarevo Dam, Herzegovina, consumption ranged from 1.5 to 1500 kgm^{-1} in different holes that had recorded only ~1.0 Lu before grouting began (Milanovic, 2000).

All springs and piezometers must be monitored carefully as the reservoir fills behind a completed dam. Operators should be prepared to halt filling and drain the reservoir as soon as serious problems appear. In extreme cases the reservoir floor and sides may be sealed off, e.g. by plastic sheeting. Experience shows that remedial measures after a dam has been completed and tested are much more costly than dense grouting during construction.

Dam/reservoir	After first filling ($m^3 s^{-1}$)	After remedial works ($m^3 s^{-1}$)
Keban (Turkey)	26	< 10
Camarassa (Spain)	11.2	2.6
Mavrovo (FYR Macedonia)	9.5	Considerably reduced
Great Falls (USA)	9.5	0.2
Marun (Iran)	10	Considerably reduced
Canelles (Spain)	8	Negligible
Slano (Yugoslavia) ($34 m^3 s^{-1}$)*	8	(3.5) Increase till 6
Ataturk (Turkey)	> 11	?
Višegrad (Bosnia)	9.4	Remedial work runs
Buško Blato (Bosnia) ($40 m^3 s^{-1}$)*	5	3
Dokan (Iraq)	6	No leakage
Contreas (Spain)	3–4	?
Hutovo (Herzegovina) ($10 m^3 s^{-1}$)*	3	1
Gorica (Herzegovina)	2–3	No remedial works
Špilje (FYR Macedonia)	2	No remedial works
El Cajon (Honduras)	1.65	0.1
Krupac (Yugoslavia)	1.4	Negligible
Charmine (France)	0.8	0.02
Kruščica (Sklope) (Croatia)	0.8	0.35
Mornos (Greece)	0.5	Considerably reduced
Piva (Yugoslavia)	0.7–1	No remedial works
Maria Cristina (Spain)	20% of inflow	?
Peruča (Croatia)	1	No remedial works
Sichar (Spain)	20% of inflow	?
La Bolera (Spain)	0.6	?

Table 4. Leakage from reservoirs reduced after remedial works (Milanovic, 2004).

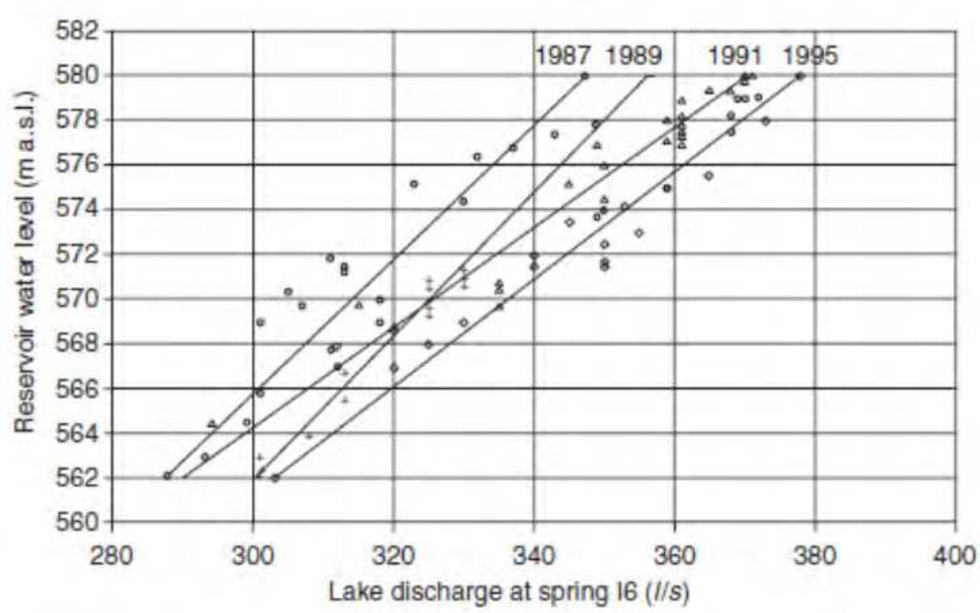


Fig. 10. Increasing leakage at Spring I6 below the Spilje Dam, Macedonia, related to water level in the reservoir behind the dam (Ford and Williams, 2007).

Despite such intense effort dams still fail to achieve design levels in karst. A good example is the Lar Dam, Elbruz Mountains, Iran. This is a 105m high earth-fill dam in a hanging valley at 2440m a.s.l. The geology is complex. The natural water table was >200m beneath the dam, draining to major springs 8 km distant and 350m lower in elevation. A sequence of international engineering firms tackled it beginning in the 1950s. During the first attempt to fill the reservoir, leakage via the springs rose to 60–80% of the inflow. It was drained and re-grouted, with 1000–40 000 kgm^{-1} of grout being injected in the worst places. A cavern of >90 000m^3 was also discovered and filled. Water losses remain unacceptably high.

5.5 Dam construction on gypsum and anhydrite

Numerous case histories are provided by A. N. James (1992) that illustrate the wide range of serious difficulties that have been encountered by building dams on evaporate rocks. These include the rapid enlargement of existing conduits and the creation of new ones, because hydraulic gradients are excessive (Figure 11); the settling or collapse of foundations or abutments where gypsum is weakened by solution; heave of foundations where anhydrite is hydrated, and attack by sulphate-rich waters upon concrete in the dam itself:

$$CaCO3 + 2H2SO4 = Ca^{2+} + SO_4^{2-} + CO2 + H2O \quad (5)$$

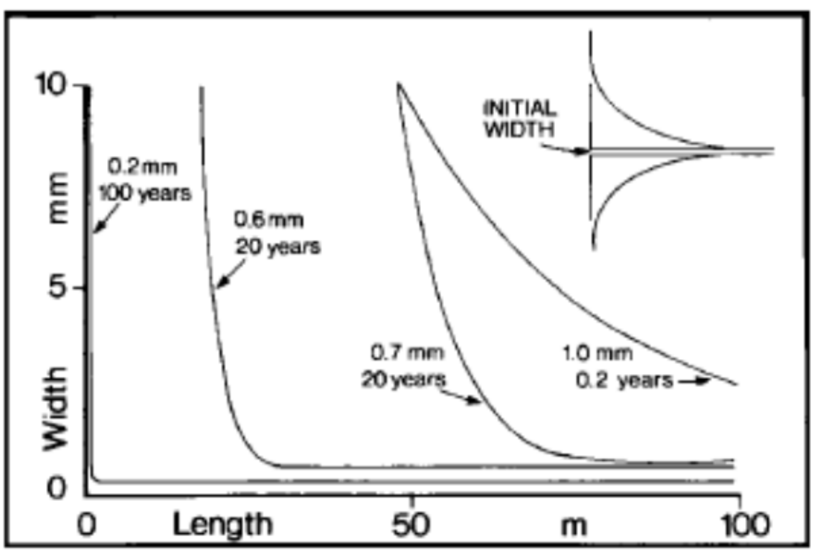


Fig. 11. Penetration distances or progress of the dissolution front for ~L99 in massive gypsum, calculated for initial fissure widths ranging 0.21–1.0 mm. Time elapsed since initiation is in years. The hydraulic gradient is 0.2 and water temperature is 10°C. (Inset) (Ford and Williams, 2007).

In their model analysis, Dreybrodt et al. (2002) obtained kinetic breakthrough beneath a 100m deep grout curtain in gypsum in 20–30 years, the conduits enlarging to give unacceptable rates of leakage within the ensuing five years.

In the USA experience has been gained in simple geological, low relief terrains in west Texas and New Mexico, where it is possible to avoid truly excessive hydraulic gradients and the problems of complex structure. At one celebrated site, McMillan Dam, gypsum is present only in the abutments, further reducing the difficulties, and no caves were detected when it was built in 1893. Nevertheless, the reservoir drained dry via caves through the left-hand abutment within 12 years. Attempts to seal off the leaking area by a cofferdam failed because new caves developed upstream of it. Between 1893 and 1942 it is estimated that 50-106m^3 of dissolution channels were created (James and Lupton, 1978).

Dams can be built successfully in gypsum terrains where relief is low and geology simple (or where there are gypsum interbeds in carbonate strata), but comprehensive grouting is necessary and an impermeable covering over all gypsum outcrops is desirable (Pechorkin, 1986). Periodic draining and regrouting will probably be needed also.

6. Water level interpretation in karst

Bonacci (1988) stressed the important role of piezometers in explaining the ground water circulation in karst. The measurements carded out on several piezometers in the Ombla catchment made further ground water analyses possible and helped to reach theoretical and practical conclusions that are important from the engineering viewpoint.

The ground water level (GWL) measurements were carried continuously in the period between January 1988 and July 1991 on 10 piezometers located in the hinterland of the Ombla Spring.

In karstified aquifers borehole data are usually difficult to relate to aquifer structure and behavior (Bakalowicz et al. 1995). However, the distribution of porosity and hydraulic conductivity from the surface to the phreatic zone can be investigated by borehole analysis and, where the number of boreholes is large, borehole tests can provide valuable information on aquifer behavior, especially in the more porous aquifers such as coral and chalk. Only rarely do boreholes intersect major active karst drains, because the areal coverage of cave passages is usually less than 1% of an aquifer and only exceptionally above 2.5% (Worthington, 1999), but when they do their hydraulic behavior may be compared to that of a spring.

More often boreholes intersect small voids with only indirect and inefficient connection to a major drainage line, in which case hydraulic behavior in the bore is very sluggish compared with that in neighboring conduits (Ford and Williams, 2007).

Bonacci and Bonacci (2000) attempted to determine the characteristics of a karst aquifer using information on ground water level (GWL) measurements in natural holes and boreholes. The majority of karst terrains in the world are still insufficiently researched from the hydrological and hydrogeological standpoint. One of the reasons is because they are situated in less developed and less wealthy parts of the world. The second but probably more important reason is the extreme heterogeneity of the karst aquifer, which causes complexity for investigation and explanation. It is very hard to obtain reliable information, parameters and general conclusions on water circulation processes. However, significant progress has been made in the investigation of karst aquifers and related problems in the last 19 years.

The differences between the four examples from the Dinaric karst given are primarily in measurement methods, i.e. the GWL monitoring procedures and in the availability and accuracy of other hydrogeological and hydrological information.

Regardless of significant differences in measurement methods, which also stipulated different elaboration levels, similar conclusions on functioning of the karst aquifer were reached. It is obvious that karst aquifers are less homogeneous than in granular media, therefore all information, especially on GWL, is valuable for their study. In karst, modelers erroneously assume that field sampling "especially from piezometers" gives an accurate spatial and time representation of the area being modeled. The result is that from the start the model is built with parameters that are not precise and have errors that accumulate in the model results. The basic problem is how to recognize and explain information contained in the GWL data. The best and probably only solution is a holistic interdisciplinary cooperation among numerous experts in the field karstology (Bonacci and Bonacci, 2000).

Bonacci (1995) interpreted the water level fluctuations in two piezometers in the Ombla spring catchment. Figure 12 presents the relationship between the simultaneous hourly GWL measured in Piezometer 8 (ordinate) and Piezometer 9 (abscissa) during a flood hydrograph at the Ombla Spring. The formation of a loop is evident during the rising and falling of the GWL, which shows that the water flow during the analysed period is non-steady. The rising period of GWL is much shorter than the falling period. The rise in GWL measured on Piezometer 9 lasts much longer, as can be seen from Figure 12, where a certain point shows the time computed from zero, i.e. the initial moment (06:00 h on 2 December 1988).

Such phenomena were recorded by BoreUi (1966) at the Bunko Blato in the Dinaric region (Bosnia and Herzegovina).

Figure 13 gives a graphical presentation of the hourly values of GWL in Piezometer 8 and the respective discharges of the Ombla Spring. GWL in Piezometer 8, as well as in other piezometers, stagnates (either remains stable or slightly increases) in the period when the hydrograph increases from 9.32 m^3/s to 54.9 m^3/s, which occurred on 8 October 1989 from 03:00 h to 14:00 h. The discharge increase at the spring lasted 11 h and in the mean time GWL in Piezometer 8 and in other piezometers was either stable or increased slowly. It should be noted that intensive rainfall started in the catchment 2 h earlier, i.e. at 01:00 h on 8 October 1989. This phenomenon can be explained by the fact that during those 11 h some large karst conduits were filled and flow under pressure was formed only in the karst conduits and caves, whereas the small karst fissures were not filled by water, so that GWL was not raised in the entire karst massif. The filling of large fissures is carried out by a rapid turbulent flow regime, whereas the small fissures are filled by a slow laminar or transitional flow regime (Bruckner at al., 1972; Gale, 1984; Lauritzen et al., 1985; Atkinson, 1986). Accordingly, it is possible to determine the volume of large caverns by integrating the hydrograph of the Ombla Spring during an 11 h period. The volume amounts to about 1.5×10^6 m^3. The analysis of the other hydrographs confirmed this value as an average value, and it was adopted for further analyses. It should be stressed here that such analyses should be performed only for those hydrographs that reach maximum discharges of the Ombla Spring over 80 m^3/s.

The piezometers, if carefully located, make it possible to identify the dimensions and functions of the karst underground system. Continuous measurements of GWL and discharge make it possible to reach conclusions on the water circulation in karst under different hydrologic conditions, i.e. at low, average and high water levels (Bonacci, 1995).

Healy and Cook (2002) review methods for estimating groundwater recharge that are based on knowledge of groundwater levels. Most of the discussion is devoted to the use of fluctuations in groundwater levels over time to estimate recharge. This approach is termed the water-table fluctuation (WTF) method and is applicable only to unconfined aquifers. In addition to monitoring of water levels in one or more wells or piezometers, an estimate of specific yield is required (Healy and Cook, 2002).

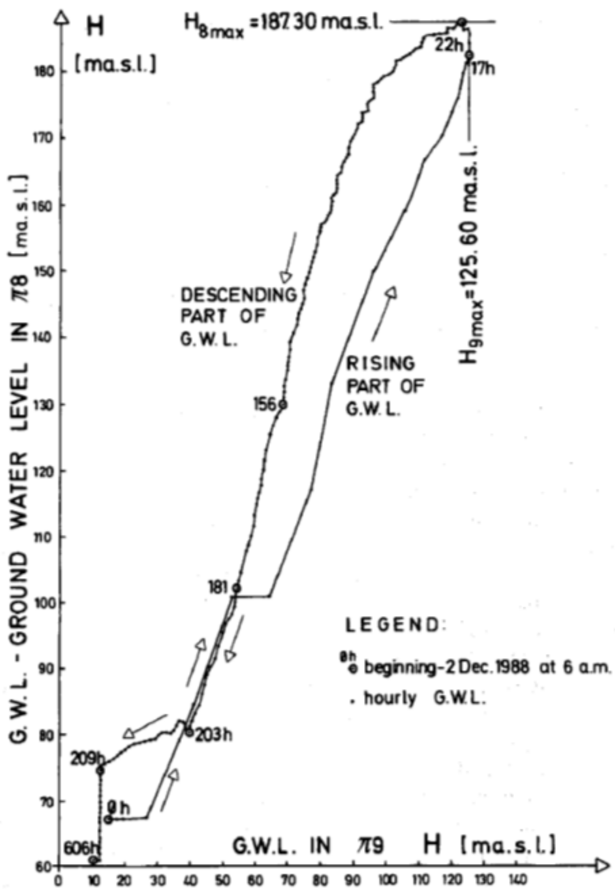


Fig. 12. Relationship between hourly GWL measured in Piezometers 8 and 9 (Bonacci, 1995).

Rehm et al. (1982) conducted a recharge study for an upland area of central North Dakota, USA. They used three different methods for estimating recharge: the water-table fluctuation method, the Hantush method, and a flow-net analysis. The 150-km^2-study area contained 175 piezometers and water-table wells. The water table is in glacial deposits of sand, gravel, and loam. These deposits overlie bedrock, silt, and clay units that confine the Hagel Lignite Bed aquifer. Thirty-eight observation wells were used to estimate recharge by the WTF method. An average value for Sy of 0.16 was obtained. Estimates of recharge to the water table were made for sand and fine textured materials. Average values are shown for 1979 and 1980 in Table 5.

Ten groups of nested piezometers were used to determine vertical hydraulic gradients for the Hantush method. Hydraulic conductivity was measured in the field at each piezometer using single-hole slug tests. The measured hydraulic conductivity was assumed to be equal to the vertical hydraulic conductivity. Sites were located in the sandy and fine-textured bedrock, as well as below two of the many sloughs that are present in the study area. Measured values of K range from 3×10^{-9} m/s for clayey bedrock to 6×10^{-6} m/s for sands. Vertical gradients range from 0.006 in the sands to 1.2 in the fine-textured material. Average rates of recharge are listed in Table 5. They are considerably greater than those estimated from the WTF method. The differences are likely caused by errors inherent in the methods and natural heterogeneities within the system. Rehm et al. (1982) calculated an areal average recharge rate for the study area of 0.025–0.115 m/year by weighting the estimates in Table 5 on the basis of area coverage of the different hydrogeological settings.

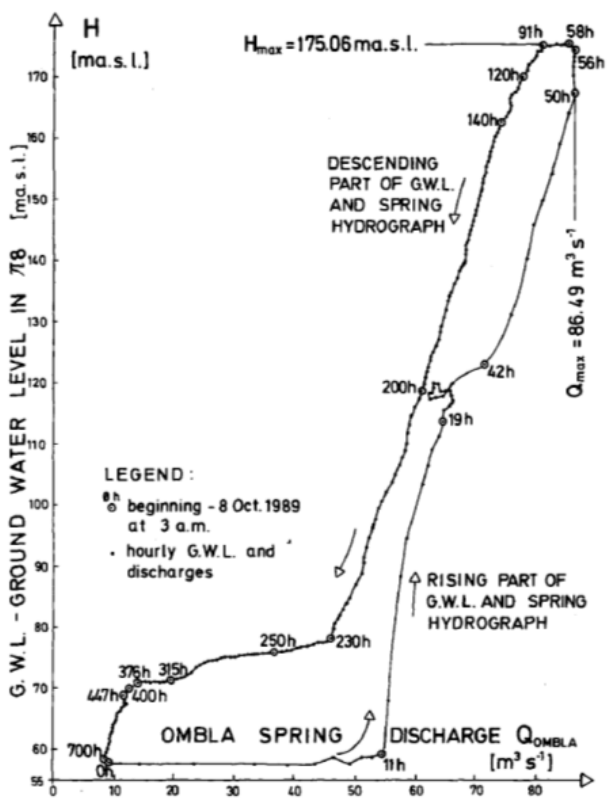


Fig. 13. Relationship between hourly Ombla Spring discharges and GWL measured in Piezometer 8 (Bonacci, 1995).

Material/location	WTF method		Hantush method	
	1979	1980	1979	1980
Sandy material	0.017	0.08	0.71	0.73
Fine-textured material	0.0018	0.0011	0.091	0.081
Sloughs	–	–	0.60	0.70

Table 5. Estimates of groundwater recharge rates, in m/year, for an upland area of North Dakota, USA, by the water-table fluctuation method (WTF) and the Hantush method for 1979 and 1980 (Rehm et al., 1982)

The basic procedure in hydraulic testing of boreholes is to inject or withdraw fluid from the hole (or a test interval within it) while measuring the hydraulic head. In the field, hydraulic conductivity is usually determined by borehole pumping tests or recharge tests (sometimes called Lugeon tests). The appropriate technique depends upon the purpose and scale of the investigation (Castany, 1984). Hydraulic conductivity may also be determined from in-hole tracer dilution (Ford and Williams, 2007).

7. Newly formed sinkholes and subsidence in karst regions

In the last two decades, by growing agricultural activities and accordingly increasing groundwater abstraction, land subsidence and sinkholes have formed in plains of different parts of the world. Various studies (Beck, 1986; Beck and Sinclair, 1986; Newton, 1987; Waltham, 1989; Benito et al., 1995; Gutierrez and Gutierrez, 1998; Tharp, 1999; Atapour and Aftabi, 2002; Salvati and Sasowsky, 2002; Abbas Nejad, 2004; Guerrero et al., 2004, 2008; Ahmadipour, 2005; Ardau et al., 2005; Gutierrez et al., 2005, 2007, 2008a; Wilson and Beck, 2005; Khorsandi and Miyata, 2007; Van Den Eeckhaut et al., 2007; Galve et al., 2008, 2009; Vigna et al., 2008; De Waele, 2009; De Waele et al., 2009; Parise et al., 2009; etc.) document problems related to sinkholes all around the world. Frequent triggering factors in the development of land subsidence and sinkholes are overexploitation from groundwater aquifers and groundwater level fluctuations. Moreover, existence of fine-grained alluvium, tectonic features especially faults and dissolution of karstic formations were reported as effective factors in the formation of sinkholes (Karimi and Taheri, 2010).

One of the most important factors in subsidence and sinkhole formation and also the rate of their formation is the water table condition and its variations. This component causes derangement in the balance between vertical stress and supporting forces in the soil column. As an example from Karimi and Taheri (2010), in order to understand the geotechnical conditions of the soil column in Famenin plain, Iran, a specific exploration borehole having 4 inches diameter, 115 m depth, rotary drilling method without drilling mud and undisturbed sampling in each 1.5 m was drilled in the distance of 15 m from Jahan Abad collapse sinkhole (6F-1). The thickness of overburden alluvium was 94 m, and the soil was mainly fine grained. The soil type from surface to the bedrock was alternation between silty clay (CL) and clayey silt (ML) with intercalations of silty and clayey sand. The soil had a medium plasticity and liquid limit ranges from 31.2 to 47.2%. The carbonate bedrock was encountered under the alluvium at a depth of 94 m, and it was composed of crashed cemented limestone fragments (Calcareous fault breccia or collapse breccia) having small voids. The void size increased at the depth range from 109 to 115 m. The water table was at a depth of 109.5 m, showing the unsaturated condition of alluvium. The borehole was sand productive, and it was implying the contribution of sand from alluvial overburden sediments.

The following factors favor the formation of sinkholes in the Famenin and Kabudar Ahang plains, Iran (Karimi and Taheri, 2010):

a. Carbonate bedrock beneath the alluvial aquifer.
b. Thick and cohesive unconsolidated soil.
c. Significant drawdown of water table due to overexploitation.
d. Penetration of deep wells into karstified limestone bedrock and predominance of turbulent flow via open joints and fractures.
e. Evacuation of alluvial aquifer materials by pumped water (sand production of wells).

Aquifer subsurface conditions and geotechnical data of 6F-1 borehole have been taken into account in order to analyze the imposed stresses into the soil column.

The assumptions of stress model on soil mass are as follows:

- Average thickness of alluvium is 80 m.
- The water table had been lowered 50 m during 10 years (1991–2001).

- The original water table depth was 30 m.
- The aquifer is unconfined and more or less homogenous.
- The soil is a mixture of silty clay with sand and gravel, having a dry density of 1.85 g/cm^3 and a saturated density of 2.1 g/cm^3.
- From water table to the bedrock, the aquifer is fully saturated.

The effective normal stress at any depth in the aquifer can be calculated with the following equation (Terzaghi et al. 1996):

$$\sigma' = (Z_1\gamma + Z_2\gamma_{Sat}) - Z_2\gamma_w \tag{6}$$

In which σ' is effective normal stress (kN/m^2), Z_1 thickness above water table, Z_2 saturated thickness below water table, γ bulk density of the aquifer, γ_{Sat} saturated density of the aquifer and γ_w density of water. The water table changes create variations in the effective stresses. The imposed stress on the cavities at the bottom of the alluvial aquifer or on the top of the karst aquifer can be calculated from the above equation. Figure 14 shows a hypothetical geotechnical model and the variations of effective stress above the limestone bedrock with the mentioned assumptions. Following the decrease of water table in the area, the changes indicated below occur in the subsurface conditions:

- Loss or removal of buoyant support.
- Increase in the effective stress at the base of the soil column, for example, the effective stress increases from 1,000 to 1,400 kN/m^2 due to the water table drawdown.
- Natural consolidation of soil causes land subsidence.

Ford and Williams (2007) inspected the sinkhole formation by dewatering, surcharging, solution mining and other practices on karst.

7.1 Induced sinkhole formation

It is probably true to write that, after groundwater pollution, induced sinkholes are the most prominent hazardous effect of human activity in karst regions. Agriculture, mining and quarrying, highways and railways, urban and industrial constructions all contribute to the effect. Induced sinkholes generally develop more rapidly than most natural ones, appearing and enlarging in time spans ranging from seconds to a few weeks. This is faster than most societies can react with preventive or damage-limiting measures, so that such sinkholes are widely described as ‘catastrophic’. Although in a few instances the hazardous collapse is of surface bedrock directly into a cave underneath, in 99% of reported cases or more it occurs in an overburden of unconsolidated cover sands, silts or clays, i.e. it is a suffusion or cover-collapse do line; ‘subsidence sinkhole’ is the widely used alternative term. There are two end-member processes:

1. ravelling, the grain-by-grain (or clump-by-clump) loss of detrital particles into an underlying karst cavity that is transmitted immediately by grain displacement upwards to the surface, where it appears as a funnel that gradually widens and deepens;
2. formation of a soil arch in more cohesive clays and silt–clay mixtures over the karst cavity that then stops upwards until it breaks through to the surface.

The large majority of suffusion sinkholes appear to have formed by a mixture of the two processes, with the soil above an early arch subsiding by down faulting, suffusion then sapping the fault-weakened mass to create a new arch and repeat the cycle. From the human perspective, pure soil-arch collapses (or 'dropouts') are the most dangerous because they can appear without warning at the surface. There has been much loss of life as a consequence.

The contact between a solution-indented karst surface and overlying cover deposits is the 'rockhead'. Using ground-penetrating radar (GPR) through 2–20m of cover, Wilson and Beck (1988) estimated that the frequency of karst cavities at the rockhead capable of swallowing ravelled debris varied from 12 000 to 730 000 km² in the counties of northern Florida, i.e. the rockhead there is a dense epikarst of solution pits, pipes and shafts.

Induced sinkhole formation thus has attracted a good deal of attention in recent decades. Hundreds of thousands of new subsidences have been reported worldwide, ranging from 1 × 1 × 1m holes to features >100m in length or diameter and tens of meters deep.

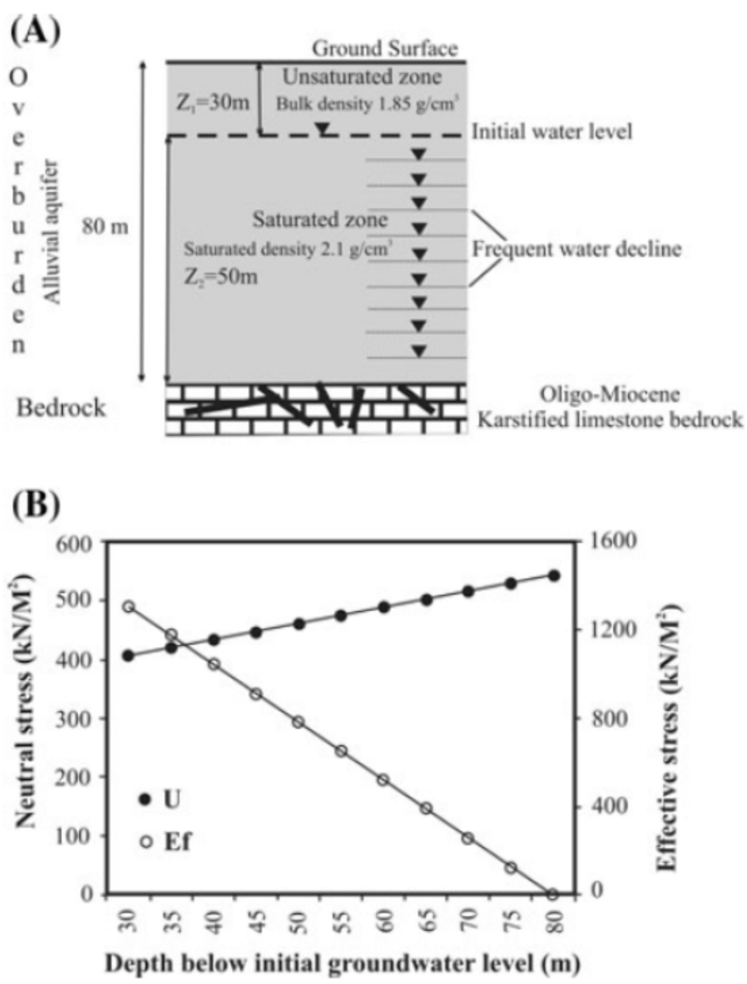


Fig. 14. A: Hypothetical geotechnical model within the soil mass, B: Relationship between effective and neutral stresses in the soil mass model (U, neutral stress; Ef, effective stress) (Karimi and Taheri, 2010).

7.2 Groundwater abstraction and dewatering

Dewatering of unconsolidated cover deposits on karst is the most important cause of induced sinkhole formation worldwide. The buoyant support of the water is removed, weakening mechanical stability. Abstraction may be for water supplies, irrigation, draining a volume of rock for mining or quarrying, or many other purposes. It is most hazardous when the cover is drained entirely so that the water table is depressed below the rockhead into the karst strata. However, sinkholes may also form readily where the lowering is limited to some level within the overburden.

Collapses are common on corrosion plains and in the bottoms of poljes when these are pumped for irrigation during dry seasons in tropical and mediterranean regions. With or without human intervention, the rapid rate of dissolution of gypsum can yield major problems. For example natural ground collapse over gypsum in northeast England resulted in about $1.5M worth of damage in the interval 1984–1994 (Cooper 1998).

Mining and quarrying tend to have the greatest impacts because they dewater to the greatest depths, usually far below the rockhead. In many coal-mining areas of China the coals are overlain by or intermingled with limestones or gypsum. Tens of thousands of sinkholes, often of large size, have been reported.

7.3 Surcharging with water

In the case of surcharging it is the addition of water at particular points that causes ravelling of overburden into karst cavities. It will therefore be especially potent where the unsurcharged water table is below the rockhead, but can also be effective with water levels in the overburden. In modern cities such point-located surcharging will be widespread unless precautions are taken, caused by drainage from individual downspouts on buildings, leaks from water supply and sewer pipes, leaking stormwater management ponds, parking lots, etc. Sinkhole formation is fastest where the overburden is thin, and is chiefly by ravelling. Sinkholes tend to be smaller than those associated with substantial dewatering, being mostly less than 10m in diameter. Nevertheless, there are many reports of building foundations being undermined and collapsed, and damage to roads and railway tracks due to neglect of soakaways or other means of dispersing stormwaters. Natural surcharging occurs on river floodplains, corrosion plains and poljes when they are inundated and is often accompanied by collapse and suffusion as the waters recede.

General rising of the water table can also create collapse or subsidence by destroying the cohesion of susceptible clay soils. However, this is comparatively rare in karst areas. The load or vibration from heavy equipment can induce small collapses locally, especially beneath it. In historic times plough-horse teams have dropped; in modern times many tractors, haulage trucks, drilling rigs and military tanks have fallen. Rock blasting from quarries or foundation cutting, etc. often causes collapses of small to intermediate scale.

7.4 Solution mining

Salt mining has induced many collapses and subsidences over the centuries. Normally this will involve significant thickness of overlying consolidated rocks. Suffusion in superficial unconsolidated deposits thus is not usually predominant, as it is in the dewatering and surcharging situations considered above. Traditionally, extraction has been by one of two methods:

1. Conventional mining via shafts and adits, removing the product by hand or machine at the workface, as in a coal mine, etc.;
2. By pumping the water from natural salt springs (‘wild brine’).

Recently, where feasible these have been replaced by solution mining, in which water is injected via one set of boreholes and extracted as brine via another, i.e. no workers or equipment are committed underground. The planimetric extent and volumes of cavities that will be created in the salts by both the wild brine and the injection methods are always uncertain and can be hazardous.

The most celebrated historic examples of induced subsidence in the English-speaking world are in the county of Cheshire, England, where wild brine mining began in Roman times and adit mining of salt became important with the beginning of the Industrial Revolution. Large areas of the surface have now subsided, over both open mines and solution mines, with much property damage, although catastrophically rapid collapse of ground is comparatively rare due to local geological conditions (Cooper 2001).

Corporations drilling exploratory and extractive oil wells do not expect to be involved in salt solution mining as well but this has happened in many instances in recent decades.

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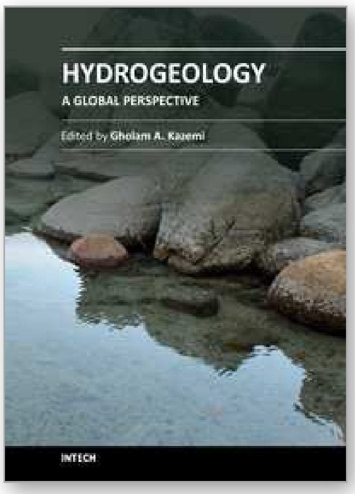

Hydrogeology - A Global Perspective

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The field of groundwater hydrology and the discipline of hydrogeology have attracted a lot of attention during the past few decades. This is mainly because of the increasing need for high quality water, especially groundwater. This book, written by 15 scientists from 6 countries, clearly demonstrates the extensive range of issues that are dealt with in the field of hydrogeology. Karst hydrogeology and deposition processes, hydrogeochemistry, soil hydraulic properties as a factor affecting groundwater recharge processes, relevant conceptual models, and geophysical exploration for groundwater are all discussed in this book, giving the reader a global perspective on what hydrogeologists and co-scientists are currently working on to better manage groundwater resources. Graduate students, as well as practitioners, will find this book a useful resource and valuable guide.

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