Karst processes and landforms

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All landscapes of the Earth are formed by a combination of physical and chemical processes that have acted on the exposed rocks over long periods of time. The different lithologies are weathered and eroded away and leave erosional morphologies, while the resulting sediments are transported by various agents (rivers, ice, wind) and deposited elsewhere, sometimes thousands of kilometers away from their source, forming depositional landscapes on land or at the bottom of the ocean. These landscapes can take their name from the dominant processes that have shaped most of the surface features (e.g., glacial or volcanic landscape) or from the prevailing rocks (e.g., sandstone or granite landscape).

The karst landscape takes its name from a region between Italy and Slovenia dominated by outcrops of carbonate rocks. Karst is the German wording, adopted during the Austro-Hungarian period when the first studies on dissolution-related landforms were conducted. It derives from the old Indo-European term "karra," meaning "stone" (Palmer 2007), known in Latin as "carusardius," and still used in the form "kras" in Slovenia and Croatia. Karst landscapes are formed mainly by surface and subsurface rock dissolution and are best developed in tropical and mid-latitude regions where carbonates and evaporites are exposed. Karst dissolution is subdued or impossible in areas with no rain (e.g., deserts) or where temperatures are always below zero (polar regions, high mountain areas). When a karst area is buried by other rocks and dissolution stops, the voids are often filled with sediments or mineralizations. Some of these can become important mineral resources, such as the Mississippi Valley Type deposits. These ancient karsts are referred to as paleokarst (Bosak *et al.* 1989). By contrast with all other landscapes, in karst chemical dissolution largely overrules mechanical erosion. Most of the rock is carried away in solution by surface and underground flowing waters and physical sediment production is very low, restricted to insoluble material, mainly clays. Because of the dominant role of dissolution over erosion, karst areas are characterized by a very distinctive morphology and hydrology (Ford and Williams 2007).

Although karst typically refers to limestone terrains, mainly composed of the mineral calcite, dissolution can also be the dominant geomorphic process in other rocks. Extensive cave systems have also been discovered in dolostone, composed of the mineral dolomite which has slower dissolution kinetics than calcite. Good examples of such cave systems, mainly developed in tectonically complex areas, are found in the Dolomites in Italy (Sauro, Zampieri, and Filipponi 2013). Gypsum and anhydrite are around 100 times more soluble than calcite. Karst in these evaporite rocks has been described in all continents and in a wide range of different climatic settings (Klimchouk *et al.* 1996; Calaforra 1998). The most soluble rock on Earth is rock salt (halite). In regions where rainfall is very low and/or rock salt is continuously brought to the surface by salt diapirism (e.g., Dead Sea area in Israel, Atacama desert in Chile, Zagros Mountains in Iran), surface dissolution can create the same

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karst morphologies encountered in limestones, but at a rate more than a thousand times faster (Frumkin 2013). Also, quartz-sandstone terrains, in tropical areas such as the tepuis in Venezuela and Brazil, can host typical karst morphologies such as solution rills, caves, and underground drainage (Wray 1997). Although referred to by some authors as "pseudokarst," there is unmistakable evidence that dissolution is an essential morphogenetic process in these landscapes.

The chemical processes responsible for the dissolution of the rocks are various and depend on the mineralogy of the rock to be dissolved (Dreybrodt 1996). In gypsum and rock salt the dissolution is a simple two-phase (solid-liquid) dissociation. The solubility of gypsum in pure water is around 2500 mg/L at 20°C, roughly 140 times lower than that of rock salt (halite, NaCl, 360 000 mg/L) and three orders of magnitude greater than that of calcite (1.5 mg/L). Dissolution of any salt increases in the presence of other salts because of ion-pairing effects. Saturation is reached rapidly and dissolution rates are boosted in the presence of turbulent flow.

In quartz-sandstone terrains, quartz (SiO_2) is dissolved forming orthosilicic acid (H_4SiO_4), reaching saturation at 8.7 mg/L in pure water at nearly neutral pH and 20°C. Amorphous silica is slightly more soluble, reaching a solubility of around 100 mg/L. Both quartz solubility and rate of dissolution increase with temperature. The dissociation of orthosilicic acid, which also increases the solubility of quartz, starts at neutral to basic pH. In these conditions the orthosilicic acid undergoes up to four consecutive dissociations (Mecchia *et al.* 2014).

But the most intensively studied karst processes are those related to carbonate rocks such as limestones and dolostones. Solubility of both calcite and dolomite in pure water at 20°C is very low, similar to the solubility of quartz, but increases by two orders of magnitude in the presence of slightly acidic waters. In most natural systems on the Earth, acidity primarily derives from epigenic sources, mainly from carbon dioxide (CO_2) present in the atmosphere and in the soil. CO_2 slowly dissolves into meteoric waters forming carbonic acid (H_2CO_3), thus reducing pH and increasing the corrosion capability of the percolating waters. Besides carbon dioxide, acidity can also derive from organic acids (humic and fulvic acids) or oxidation processes occurring in aerate conditions. These aggressive waters penetrate into the bedrock through available permeable pathways (fractures, bedding planes) and flow down-gradient in the saturated (phreatic) zone towards the discharge points (i.e., springs). In carbonate karst areas, most of the dissolution occurs close to the surface, in the first tens of meters of rock, known as epikarst, and rapidly decreases downwards as the saturation degree increases.

Many factors influence the chemical reactions involved in the dissolution of carbonate rocks, the most important of which is mixing of different waters. In the mixing zone between two different waters, or at the interface between fresh and saline groundwater, typical in coastal karst areas, chemical dissolution is enhanced. Mixing of two solutions saturated with respect to calcite but at different partial carbon dioxide pressures results in an undersaturated solution capable of dissolving more limestone. This mixing effect, known in the literature as the "Bögli effect," makes dissolution possible also deep in the aquifer. In young limestones close to the sea (e.g., small carbonate islands), enhanced dissolution at the interface between the freshwater body and the denser salt water gives rise to the so-called flank margin caves (Mylroie and Carew 1990).

The acidity of the solutions can also derive from deep (hypogenic) sources, in the form of CO_2- or H_2S-rich rising fluids. These sometimes highly aggressive rising waters lead to the formation of the so-called hypogenic cave systems (Klimchouk 2007), some of which are created by fluids enriched in carbonic acid (e.g., Black Hills, South Dakota), and others by solutions enriched in sulfuric acid (e.g., Guadalupe Mountains, New Mexico). Since the source of acidity derives from deep underground and is brought to the surface along major discontinuities (e.g., deep faults), these systems are often found in areas with thermal springs (e.g., Budapest). The dissolving power of the fluids diminishes moving upward, often leaving no geomorphic expression on the land surface. Hypogenic caves are also present in gypsum areas, where undersaturated waters enter the gypsum beds from below in an interstratal karst setting. These fluids dissolve the gypsum beds along every possible initial open fracture, leading to the formation of sometimes impressive regular maze caves (e.g., Ukraine) (Klimchouk 2007). Some of the largest cave systems in the world are actually hypogenic caves, such as Lechuguilla Cave in New Mexico (a 222-km-long sulfuric acid cave in limestone) or Optymistychna Cave in Ukraine (a 236-km-long gypsum maze).

In general the word "karst" refers to a landscape dominated by unique morphologies that form because of surface dissolution of the host rock. This process creates a wide variety of small-scale landforms typical of karst: these are called karren. They occur on all soluble rocks, especially limestone, gypsum, and rock salt, but also on quartz-arenites. They vary in form from small pits and grooves, to rills (rillen), to runnels (rinnenkarren), and flat-floored solution pans (kamenitze). These centimeter- to meter-sized rugged landforms are typical of high mountain karst (Figure 1) in the absence of soil and cover. Karren are also very well developed in coastal areas, where the corrosive, burrowing, and etching actions of organisms exacerbate their shape. In this environment bioerosion becomes dominant, and the landscape is often called biokarst or phytokarst.

Karst areas often have a drainage characterized by closed depressions and generally lack normal river networks. These closed depressions are probably the most distinctive landform of karst, and are known as dolines, from the Slavic word "dol," meaning "valley." In North America and in engineering language dolines are called sinkholes. The presence of dolines in an area is diagnostic of the dominance of dissolutional processes over erosional ones. Dolines can have many shapes, from shallow saucer-shaped depressions to large vertical shafts, depending on the dominant processes that formed them: dissolution, suffosion, collapse, or sagging. They can have diameters ranging from a couple of meters to over a kilometer and normally have a subcircular plan form. When numerous dolines coalesce they form complex depressions historically named uvalas, a term that is falling into disuse. Collapsed dolines that give access to the underlying aquifer are known as cenotes, typical in Central American coastal areas. Dolines occur both in carbonate and in evaporite rocks, with a greater frequency and dimension in the latter mainly because of their higher solubility (100 times more than the carbonate rocks), and their lower mechanical strength and higher ductility (Gutiérrez *et al.* 2014). Generally, dolines are classified as either solution dolines or subsidence dolines. The first develop because of focused dissolution in zones of higher permeability, resulting in the gradual and differential lowering of the ground surface. In general, this type of doline forms in areas with exposed or barely soil-covered soluble rocks. Subsidence dolines, on the other hand, form because of subsurface dissolution and gravitational downward movement of the overlying material. These are classified according to the type of material involved in the lowering (cover, bedrock, or caprock) and the main subsidence

Figure 1 Well-developed karren field in pure microcrystalline limestones at Tsanfleuron, Bernese Alps, Switzerland. Photo Jean-Yves Bigot. Reproduced with permission.

process (collapse, suffosion, or sagging) (Gutiérrez *et al*. 2014). Cover refers to overlying material composed of unconsolidated deposits (e.g., soils, alluvium), bedrock is the soluble rock itself (e.g., limestone or gypsum), and caprock is a covering solid nonkarstic rock. The downward movement of the material can involve brittle deformation (collapse), slow washing away of loose deposits into underlying subsurface voids and settling of the overlying deposits (suffosion), or slow ductile bending of sediments (sagging). Cover collapse sinkholes (or dolines) are the most spectacular and dangerous ones. Sometimes more than one type of material and several processes can cause the formation of a complex doline, which will be classified using combinations of the above mentioned terms, putting the dominant material and/or process first, followed by the secondary one.

Dolines can occur as isolated individuals, in groups following the direction of fractured areas or the allogenic recharge boundary of the karst area, or can occupy all of the available surface area. This egg-box-shaped pattern of dolines is known as polygonal karst, and occurs in tropical and temperate areas characterized by high values of rainfall (e.g., Papua New Guinea, New Zealand). The large density of dolines in these areas creates an extremely efficient autogenic drainage pattern, conveying surface waters very rapidly underground. The size of the dolines in these areas depends on a combination of lithological and structural factors, as well as climatic ones.

Another typical landform that derives mainly from dissolution of soluble rocks, in particular, limestones, is the karst plain, often referred to by the Slavic word "polje," meaning "large plain." These almost perfectly planed surfaces cut across the rock structure and their floor corresponds to the mean level of the water table. This means they flood during the rainy season, and remain dry for the rest of the year. Their floor is covered with a very thin cover of residual (insoluble) material with bedrock cropping out locally. Karst planation surfaces generally develop at the inflow and outflow boundaries and can be considered as the final stage of karst denudation in carbonate rocks. Often these large depressions are controlled by structural elements, such as grabens or large synclines. The soluble rock surface is lowered vertically until the water table is reached, after which dissolution enlarges these depressions laterally at the level of the water table. This produces a planation surface inclined parallel to the hydraulic gradient. Ford and Williams (2007) distinguish three main types of poljes: border poljes, structural poljes, and base-level poljes. The best examples of these poljes can be seen in the Dinaric karst.

Often the karst denudation is not complete, and some parts of the original bedrock are left standing out on the planation surfaces. Small hills like this are called hums in the Dinaric karst and are frequent in temperate areas. In a tropical climate, residual hills can instead be 200 m high and have steep to almost vertical sides. In China, isolated hills like this are called fenglin (peak forest), while clusters of residual hills sharing a common exposed bedrock base are known as fengcong (peak cluster). The first are formed close to the water table level (and thus erosional base level), while the latter are formed far above this level (and therefore lack the typical flat plain in between). The Spanish-derived term for the towers is "mogotes." The intervening karst plains are formed in a similar way to the aforementioned poljes, as horizontal swamp notches at the base of the towers confirm. The presence of several of these notches along the tower walls reflects the changes in local water table level.

Surface rivers are absent in typical karst areas, and the drainage pattern is mainly characterized by dolines. Small allogenic rivers, for which the drainage basin is partially located outside of soluble rocks, normally disappear underground soon after they enter the karst area, forming blind valleys (valleys which end "blindly"). When allogenic streams are greater, water will flow on the karst surface forming a true valley, sinking along its path into more permeable areas (losing streams). When large amounts of water are conveyed underground through point sinks these are called swallow holes (also named swallets or ponors). The discharge carrying capacity of these swallow holes often does not allow all of the water to disappear, and the river continues downstream until all the water is absorbed by the karst rocks. Downstream of this last swallet the valley becomes dry for most of the year, and activates only during high discharge periods. Some rivers may maintain their flow throughout the entire karst area until flowing out of it. This may happen because the flow rate exceeds the capacity of the karst to absorb, or because the hydraulic gradient is very low, not allowing the water to penetrate underground rapidly enough. High hydraulic gradients in fact enhance the vertical percolation of surface waters. The river valleys in karst areas are generally characterized by steep walls, typically forming canyons (or gorges).

Many of these canyons are formed gradually in a slowly uplifting region (antecedent valleys).

Since primary porosity in both limestones and evaporites is very low, surface waters in karst tend to penetrate underground through all available fractures. Passing through these variable pathways, dissolution continues and permeability gradually increases, resulting in an enhanced underground water flow and a consequent local groundwater lowering. Thus an aerate (or vadose) zone is created in the upper part of the karst aquifer, where water percolates downward or is held in place by capillary forces. This upper part of the vadose zone, extending immediately below the karst surface, is known as epikarst (or the subcutaneous zone), and can be regarded as the extension of the surface karren forms. This is where most dissolution of rock takes place in carbonates, since CO_2 concentrations are higher there. This network of fissures enlarged by dissolution often contains infillings of soil that can retain moisture over longer periods of time, gradually releasing their water downward by gravity. This subcutaneous habitat is also extremely interesting from a biological point of view. Epikarst usually ranges from a couple to some tens of meters deep. Below this epikarst, water flow is focused along a few larger voids that can extend very deep into the bedrock and give rise to cave systems, or are feeders of active caves below.

Dissolution processes, together with mechanical erosion along underground pathways, give rise to the formation of three-dimensional systems of conduits, and solutionally enlarged discontinuity planes (fractures, bedding planes), forming extremely complex three-component permeability aquifer systems. The direct connection between the surface and the karst aquifer causes underground waters to be extremely vulnerable to pollution. The high secondary permeability of karst enables waters and pollutants to be transported quickly and often without undergoing appreciable chemical and physical changes from their input points to the springs (Goldscheider and Drew 2007). Due to this rapid transfer, flow rates can also change by three orders of magnitude in a relatively short time span. During floods, pollutants stored underground during low flow conditions can be flushed out. Karst aquifers drain toward springs, which often have considerable flow rates. The largest springs in the world are most probably karstic ones, reaching flow rates of over $100\,m^3/s$. The location of these springs is often controlled by the regional or local base level (e.g., sea level), or by a less permeable threshold (e.g., underlying or adjacent rocks, alluvium), sometimes caused by tectonic features (faults). Since sea level changed during the Quaternary, some springs can be located close to or even below sea level, giving access to underwater cave systems. These springs, typical of carbonate islands but also present in other geographical contexts, are known as blue holes. Headward retreat of the spring can produce a pocket valley with the spring at its upstream end, typically at the foot of a high limestone cliff. Base-level rise, due to valley aggradation, can cause vertical vadose passages to be flooded creating very deep springs, known as Vauclusian springs, named after the Fontaine de Vaucluse (France). Some springs become active only during floods, and are located above the regional water table. A special type of temporary springs is the so-called estavelles, which act as swallow holes during normal (low) flow, and become active as springs during floods.

Cave genesis (or speleogenesis) is a complex topic, involving different sciences, mainly geology, chemistry, and physics. Dissolution is an important process, but is generally not responsible for most of the created voids, since once mechanical erosion starts, this process largely overrules chemical dissolution. Dissolution, however, is fundamental, especially in the initial phreatic phases of cave formation, when laminar flow is the only possible mechanism of water movement through the tiny little fractures and openings in the rock mass. This initiation phase in cave formation can last some thousands of years in limestone, and some decennia in gypsum. Once the width of the openings reaches a critical threshold, a little less than a centimeter, turbulent flow is established and flow rate increases. This allows the water present in these enlarged openings to remain chemically aggressive over longer pathways, enlarging this flowpath more than the neighboring ones. This causes, in the short term, the selection of this path as the main one, focusing most of the water flow through it, boosting its widening even more. This rapidly brings to the formation of a true cave passage, in a period similar to the initiation phase. These passages eventually will become air-filled and can accommodate sediments and speleothems, or continue growing through collapses. Complete infilling or surface dissection, with the creation of so-called unroofed caves, finally ends this cave's cycle.

The overall cave pattern mainly depends on the origin of the dissolving waters, on the type of recharge, and on the nature of the guiding structures (Palmer 1991). In general, cave patterns depend on the local geological settings, including the distribution and type of soluble rocks, their structural setting, and the relative (and variable) position of recharge areas and points and the outlets (generally springs). Caves develop very rapidly (in a couple of years) in rock salt outcrops, in a matter of a hundred years in gypsum, and in some thousands of years in limestone and dolostone. The initial phases of speleogenesis will tend to favor the most permeable pathways, mainly consisting in fractures and bedding planes. The initial geological and structural conditions are often reflected in the morphology of the cave passage as a whole and on the cave wall sculpturing features. Differential corrosion-erosion creates typical cross-sections along more soluble beds, weaker bedding planes, and fractures.

As described above, the aggressiveness of the waters can have a surface origin (CO_2 in the soils and the atmosphere) or a deep origin. The first waters create so-called epigenic cave systems, the latter hypogenic caves. Recharge in epigenic caves can be allogenic (from outside the karst area) and autogenic (from inside). Allogenic input often is concentrated, while autogenic input can be either concentrated or diffuse. Recharge can also be indirect, through overlying nonkarstic aquifers. In general, epigenic caves form in unconfined conditions, with water flowing from a recharge area into the karst and then flowing by gravity following the hydraulic gradient to the outputs, in general important springs. Most of the upstream cave volume is produced in vadose settings, epiphreatic (floodwater) conditions prevail close to the water table, while in the downstream end phreatic conditions occur more frequently. Typical vadose passages are shafts and canyons, along which the waters tend to descend along the steepest available pathway, often represented by the dip of the beds. Once the water table is reached, the water no longer has a need to descend rapidly, and will follow the hydraulic gradient choosing the most efficient openings. Cave passages are often completely filled with water and tend to dissolve and erode the bedrock away in a homogeneous manner. This creates phreatic tubes that are roughly cylindrical in shape and, although they loop in both horizontal and vertical directions, they often more or less follow the strike.

The history of a cave can often be reconstructed based on the careful observation of the overall morphology of the system, but especially through the analysis of rock wall sculpturing and its relation with the hosted sediments. Flow direction, for instance, can be inferred from the shape of scallops, erosional cusps left by turbulent flowing waters. Also the size of these scallops allows the flow velocity at the time of their formation to be estimated: the bigger they are, the slower the water was flowing. This ultimately allows flow rates in now abandoned conduits to be estimated based on scallop size and the passage's cross-section. Estimation of conduit cross-section size can be made difficult where sediments were present at the time of scallop formation. Underground rivers, in fact, can accumulate sediments on their floors, which shield the lower portion of the passage from corrosion-erosion. The water will thus be able to cut into the walls creating alluvial notches, slightly undulating subhorizontal lateral grooves on the walls. If alluviation fills the passage almost completely, phreatic conditions will be created and the water will corrode and erode the roof only. If sedimentation continues synchronously with this upward dissolution, ceiling channels or an anti-gravitative canyon can develop. Sediments can later be washed out from these passages leaving a canyon-like passage. This anti-gravitative erosional process is known as paragenesis, not to be confused with the minero-petrographical term. Paragenesis in caves can be related to increased sediment input into the system, which can be induced by climate changes or can be caused by anthropogenic activities (e.g., deforestation).

The most common big epigenic caves are the typical underground rivers, and their genesis is strictly related to what happens at the surface. Developing over thousands of years, they reflect both climate and landscape changes, and their speleogenetic study can help unravel the paleogeography and paleoclimatology of the area in which they have formed. Sinking points can shift upstream, creating new cave tributaries, while springs can move to lower elevations, following base-level lowering (e.g., valley entrenchment). This can create a set of cave "levels" that can be correlated with surface morphologies such as fluvial terraces.

In hypogenic caves, recharge comes from below and dissolves the rock upward. These cave systems are often located in regional groundwater systems where structural reasons cause fluids to rise (e.g., along major deep faults) and, when these waters come in contact with soluble rocks, create underground dissolution voids. In the typical intrastratal systems, the soluble bed is sandwiched between a lower-lying and an upper aquifer bed. Hypogenic caves in gypsum do not require acid waters, and caves can form as long as the rising fluids are undersaturated with respect to gypsum. This ascending speleogenesis acts along any available discontinuity and normally creates maze cave systems, some of which can reach kilometric development (e.g., Ukraine). Hypogenic caves in limestone, by contrast, need acidic waters able to dissolve the calcite (or dolomite). Since the waters often come from deep below the surface, the geothermal gradient causes them to be thermal. The decrease in pressure, due to rising, causes CO_2 to be released from the waters, while, conversely, part of this gas dissolves in the water again because of the decreasing temperatures, making these fluids become acid. Important thermal cave systems have been described in many parts of the world, such as in Budapest (Hungary) and the Black Hills (United States). In other cases the rising fluids can be enriched in H_2S, often deriving from the reduction of sulfates such as evaporitic rocks or hydrocarbon reservoirs. This gas, when arriving in an oxidizing environment, such as a cave or the shallow phreatic environment, turns into sulfuric acid that reacts with the limestone forming gypsum and releasing CO_2, which further enhances the dissolution of more limestone (Egemeier 1981).

Another peculiar type of cave is the flank margin cave, typically formed in relatively young carbonates close to the present or past sea level. In this zone freshwater–seawater mixing creates an area of enhanced dissolution forming a set of interconnected voids more or less parallel to the coastline and along a horizontal plane, corresponding to the sea level at the time of their formation (Mylroie and Carew 1990). Flank margin caves lack morphologies and sediments typical of underground streams. Cave passages tend to pinch out moving away from the mixing zone (and hence coastline). This type of cave is typical of carbonate islands, but has also been described in continental coastal limestone areas.

Once a cave has formed, it starts accumulating material, including clastic sediments, chemical precipitates, and organic debris. Cave entrances are often characterized by pitfalls, or sinks, and tend to collect surface material that falls in accidentally (animals) or is washed in by surface runoff. This creates preferential sedimentation sites of sometimes very valuable deposits that can be preserved over very long periods of time, while surface deposits are being stripped away by erosion. Caves can also be used by animals, including humans, as shelters or living places, or for ritual reasons. Caves have delivered some of the most important archeological sites for the reconstruction of the evolution of humans and their art (e.g., the Sterkfontein caves in South Africa with their *Australopithecus* finds, or the Lascaux cave in France for its wonderful Upper Paleolithic rock art).

Cave sediments can help to reveal at least part of the cave's history, but because of extensive reworking and redeposition, they are often very difficult to study (Sasowsky and Mylroie 2004). Alternations of fine and coarse sediments in relatively undisturbed sedimentary sequences in caves reveal changes in flow in the underground stream. The mineralogical and petrographical composition of sediments can tell where they come from, and reveal possible changes in recharge areas. Sediments can sometimes be dated using paleomagnetism, in combination with U–Th dating of flowstone beds (Zupan Hajna *et al.* 2008). Allogenic quartz grains can be chronologically constrained using cosmogenic Al–Be dating techniques (Granger and Muzikar 2001). All these dating techniques deliver ages younger than the cave itself.

Inflowing and percolating waters introduce not only detrital particles but also dissolved species. Reactions between these fluids, the host rock, the sediments in the cave, and the atmosphere cause minerals to precipitate as single crystals or, more commonly, as aggregates known by the general name of speleothems. Over 330 different species of minerals have been described forming in caves (Onac and Forti 2011). Mechanisms of mineral formation include a variety of processes, such as CO_2 loss, evaporation, alteration of material already present in the cave, oxidation–reduction, cooling–heating, and changes in pH, just to mention the most important ones. The most frequent minerals are of course calcite and aragonite, followed by gypsum and other carbonates and sulfates. Where bat guano or bones are present, a variety of phosphates can be found. Some minerals are intimately related to the cave's genesis, such as in the case of sulfuric acid speleogenesis, and their dating can give clues on cave formation (e.g., K–Ar and Ar/Ar dating of alunite) (Polyak *et al.* 1998).

But caves are especially appreciated for their hosted speleothems (Figure 2), such as the famous stalactites and stalagmites. These can be classified according to the nature of the water that deposited them (Hill and Forti 1997), or to the way single crystals and their aggregates

Figure 2 Pure white calcite helictite in the Su Bentu Cave, Sardinia, Italy. Photo Vittorio Crobu. Reproduced with permission.

form the external shape of these cave formations (ontogeny, Self and Hill 2003). Most speleothems are formed of calcite or aragonite. Among the most common forms are: stalactites, stalagmites, columns, flowstones, soda straws, helictites, cave pearls, and rimstone dams. A review of most of these is given in Hill and Forti (1997).

Speleothems are important climate and environmental archives, characterized by annual layering of chemical origin, allowing the reconstruction of environmental and climatic changes at high temporal resolution (Fairchild and Baker 2012). The advantage of speleothems compared to other continental archives is the fact that they are datable with great precision with techniques based on the radioactive decay of uranium. Thanks to the introduction of MC-ICP-MS (Multicollector Inductively Coupled Plasma Mass Spectrometry) it is possible to date small (less than 0.2 g) samples of carbonate with high precision (2 sigma) of ±1% in the interval between 0 and 500 000 years BP. In recent years, the research on speleothems has focused on the isotopic composition (oxygen and carbon) of the carbonate of which they are composed and on their axial growth rates in order to reconstruct an annual and secular scale of climate variability, and to recognize the effect of internal and external climate forces on global climate changes. Recently, the study of the variability in speleothem trace element concentrations (Ba, Mg, Mn, P, Sr, U, etc.) through time has emerged as a useful hydrogeological marker that allows researchers to reconstruct rainfall, from seasonal scale to millennial scales, and to obtain information on the residence time of the water in the aquifer (often related to climate), with important consequences for the extension of phenomena of water-rock interaction. Speleothems have also been used for the reconstruction of Quaternary sea level changes, hydrological conditions in continental and marine environments, paleo-seismicity during early Quaternary, and relationships between climate change and archeological events.

Many caves around the world have become tourist attractions because of their aesthetic value. A fee is charged to gain access to these caves, and normally visits are carried out following equipped pathways and under the surveillance of a cave guide. Although the first real show cave is Vilenica in Slovenia (where invited people have paid to visit the cave since 1633), most show caves developed in the second part of the twentieth century. The development of a show cave often requires modifications to be made to the cave system, such as construction of pathways, opening of new or more easy access, installation of electric lighting systems, and a whole set of infrastructures both in the cave and above it or around its entrance. These operations often cause damage to the cave and the surrounding landscape, and environmental impact assessments should be carried out prior to show cave preparation to minimize these negative impacts (Cigna 2012). Special care should be taken in the selection of paths, lights and their position, and in the timing of the visits, in order not to exceed the so-called visitor carrying capacity, the maximum number of visitors that can enter the cave in a certain time frame without changing the environmental parameters irreversibly. In other words, modifications to the underground environment (e.g., temperature rise, CO_2) should be recovered by the system at least over night, with return to natural conditions.

The protection of the cave and its environment is important to safeguard the mineral beauty of the cave (i.e., speleothems), but also to create as little disturbance as possible to the sometimes extremely fragile cave habitat. The underground voids, in fact, including not only caves but also the tiny fractures, are a habitat to an incredibly various and more or less specialized cave fauna (Culver and Pipan 2009). Some species are completely adapted to life underground: they have lost their pigments and eyes, have developed their other senses extremely well, and spend their entire life cycle in the dark and wet environment of a cave. These species, called troglobites, would not be able to survive outside and can be endemic to very small areas or even single caves. Troglophiles (cave-loving species) are species spending part of or their entire life cycle underground, being adapted to this environment, but can also be found outside (under rocks). Trogloxenes live close to the cave, and use caves as shelters (e.g., bats).

Even the slightest changes in cave environmental parameters can cause troglobites to migrate, and in the worst cases to succumb, resulting in loss of biodiversity. Modifications also include those that happen outside of the cave, at the surface or in the drainage area (water pollution, deforestation and soil loss, urbanization). While the extinction of a troglobite can pass unnoticed, the decrease in the population of large bat colonies in a cave can have a dramatic impact on the surrounding area, since bats are insectivores and therefore reduce the need for pesticides. Subterranean karst environments are also unique microbiological habitats. This biodiversity is worth protecting, and ongoing research on new cave-dwelling species and microbiological communities might allow the discovery of new substances useful for medical purposes (Barton and Northup 2007).

In the past decennia, public awareness of the unique character and the fragility of karst areas has increased enormously, especially in Europe and America, but still needs to be stressed in many countries. It is now understood that if we want to safeguard this fragile environment, and enjoy its multiple resources for centuries to come, we need to find a way of living on karst.

SEE ALSO: Applied geomorphology; Aquifers; Biodiversity; Coastal erosion processes and landforms; Earth system science; Exploration; Landforms and physiography; Natural hazards and disasters; Paleoclimatology; Quaternary geomorphology and landscapes

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