Chapter 5
Characteristics and Genesis of Subsurface Features in Glaciokarst Terrains

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Abstract Glaciokarst terrains are rich not only in specific landforms, but in subsurface forms as well. Long, complex cave systems are widespread in glaciokarst terrains, and the deepest caves are almost all found in glaciokarsts. On the other hand, as for the volume of cave chambers and passage dimensions, glaciokarst caves are not among the largest ones. One of the most important questions about glaciokarst speleogenesis is whether subglacial cave development exists at all, and if so, how effective it is. Other important issues are the age of glaciokarst caves and the karst hydrology of glaciokarst terrains. Characteristic features of alpine caves are vadose shafts and (sub)horizontal passage levels. The two main variations of passage profiles are the tubular phreatic and the canyon-like vadose cross-sections, moreover, the combination of the previous two also exists, it is the so-called keyhole profile. Among small-scale cave features, paragenetic shapes and scallops are presented in this chapter. Characteristic glaciokarst cave sediments are coarse debris, which are mainly the results of extreme high discharges, fine-grained varved carbonates, which are deposited due to back-flooding conditions, and speleothems, which grow mostly during warm periods, but if some special conditions are satisfied, they may grow even below actually glacier-covered terrains due to the so-called "common-ion effect". Further on, cryogenic cave calcites are also formed in glaciokarst caves, but their amount is insignificant. As for the karst hydrology, extreme fluctuations are characteristic to glaciokarsts, meaning both high seasonal changes and relatively high daily changes according to melt cycles. Using U-series and cosmogenic nuclide methodology to date speleothems and detrital cave sediments, it is now evident that the majority of glaciokarst caves are polygenetic in origin, surviving one or more glacial periods. Preglacial caves (i.e. caves evolving since at least the Pliocene) are common in the Alps. On the other hand, there are approved postglacial caves as well, which are related to drumlins or isostatic fissures. Finally, subglacial speleogenesis is also proved to be possible, though it has a low rate. Ice-contact cave development takes place when a connected aquifer is formed in the glacier ice and in the neighbouring karstic rock mass.

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5.1 Introduction

Most of the glaciokarst terrains are rich in subsurface forms, too. Speleogenesis is influenced by a number of factors. The first type of factors, notably lithology and tectonic settings are independent of climatic conditions and glaciations, therefore, they are called *passive factors* (Skoglund and Lauritzen 2011). On the other hand, hydrology and geochemistry are tightly related to climatic settings and glaciations, thus they are called *active factors* (Skoglund and Lauritzen 2011). Due to the active factors, speleogenesis in glaciokarst terrains has some specific characteristics. The evolution of caves in glaciokarst settings can be considered a kind of competition between karstic and other (glacial or periglacial) processes, such as frost shattering. "Cold conditions, seasonally abundant water, rapid runoff, steep gradients strongly sustain other geomorphic processes, most of which can be viewed as competing with karst development" as Smart (2004) summarized it. The impact of glaciation on karstification and consequently on speleogenesis may be deranging, destructive, inhibitive, preservative or stimulative (Ford and Williams 2007).

Speleogenesis is possible if dissolution operates faster than erosion (Palmer 2003), which, in turn, is at a very fast pace in glacial and periglacial environments. The inception phase of carbonate cave development requires generally 10^4–10^5 years, it is the time necessary for the slow dissolutional enlargement of the initial openings to become wide enough to throughput relatively high discharge of water. This threshold time is called *breakthrough time* (Palmer 1991, 2003). Most frequently, cave development leads to branchwork cave pattern, but in certain cases, maze type caves are formed. This latter type is characteristic, if the hydraulic gradient is steep and the cave is occasionally flooded, or if the subsurface flow paths are short, or if the water recharge is uniform, or if waters of different chemistry are mixed (Palmer 1991, 2003). Cave passage horizons are often developed, they are related to erosion base levels, but the relation may be quite complicated in certain cases and influenced by a variety of factors (Ford 1971; Palmer 1991, 2003; Audra 1994). Caves near the surface are subject to destruction, partly due to collapses, partly due to direct surface erosion (Ford 1983a; Audra et al. 2002; Palmer 2003). All of these general speleogenetic processes can be more or less influenced by glaciations as it is discussed in the followings.

Physical and scientific exploration of caves in glaciokarst terrains has been presented in Chap. 1. Scientific questions related to speleogenesis in glaciokarst environments have emerged in parallel with explorations and expeditions. Many of these questions have been answered by subsequent karst research, but still, there are certain details in general or in local that remained unanswered. This chapter aims at presenting our knowledge about speleogenesis in glaciokarst settings by going through the following questions:

- What are the sizes typical of subsurface glaciokarst features?
- Is speleogenesis possible below glaciers, and if so, how effective is it?
- What is the age of caves found in terrains glaciated during some periods of Quaternary, are they of preglacial, interglacial, subglacial or deglacial origin?
- What is the relationship of speleothem growth and climate? Is active speleothem growth possible in caves under actual glacier cover?
- How partly of fully glaciated karsts work from a hydrological point of view?

5.2 Morphology and Sediments of Glaciokarst Caves

5.2.1 Dimensions and Morphology

A basic fact is that caves are abundant in glaciokarst terrains. In case of alpine mountains, it was essentially unambiguous from the beginning of karst research, but as for the karst development of glacier-covered and permafrost terrains, they remained unexplored for a long time, and it was not obvious whether significant caves exist in these environments or not, and if so, how high the intensity of karst processes is. Due to the later exploration of these hardly accessible glaciokarsts, it became clear that karst processes and cave development are active even in arctic glaciokarsts (Brook and Ford 1980; Ford 1983a; Salvigsen and Elgersma 1985; Lauritzen 2006).

The caves of alpine glaciokarsts have *variegated dimensions*. As for their length, there are really remarkable long systems, such as Hölloch (203 km) and Siebenhengste (157 km) in the Swiss Alps, the Schönberg-Höhlensystem (143 km) in Totes Gebirge (Eastern Alps, Austria), the Sistema del Mortillano (139 km) in the Cantabrian Mountains (Spain), or the Gouffre de la Pierre Saint Martin (80 km) in the Pyrenees (France). In the karst terrains of the United Kingdom that were covered formerly by continental ice sheet, there are also considerable systems, though they are somewhat shorter than the previously mentioned examples. Notably, the Three Counties System (86 km) at the edge of Yorkshire or the Ogof Draenen (70 km) in South Wales. In more arctic terrains, there are no such long caves, but still, there are caves longer than 20 km, for instance, in Norway (Tjoarvekrajgge 25 km) and in Canada (Castleguard Cave 20 km) as well. Given the above data, it is evident that if only length is considered, glaciokarst caves are well developed and are among the longest cave systems in the world. Nevertheless, it is not true for the 3D size of passages and rooms. The largest cave room in glaciokarst terrains is the Salle de la Verna found in the Gouffre de la Pierre Saint Martin, but even this large feature is only 13th in the list of the largest underground chambers by volume, whereas mostly tropical and subtropical caves are found in the top twelve (www.caverbob.com). Nonetheless, aside from the records, it is stated that glaciokarst caves in general have smaller passage diameters, especially in case of arctic terrains (Lauritzen and Skoglund 2013).

One of the most remarkable features of alpine caves is that they often consist of *vadose shafts* (Fig. 5.1) and roughly *horizontal passage levels* (Smart 2004; Audra 2004; Szabó 2008, 2009; Plan et al. 2009). Naturally, this morphology is only partly related to glaciations because it is also the result of large topographic relief, tectonic uplift and deep incision (explained later in details). The deepest caves on the Earth are almost exclusively found in glaciokarst terrains, namely, in the Western Caucasus, where the Arabika Massif hosts the four deepest caves: Veryovkina (−2204 m), Krubera (−2197 m), Sarma (−1830 m), and Illuzia-Snezhnaja-Mezhonnogo (−1853 m). The Lamprechtsofen-Vogelschacht (−1632 m) is found in the Eastern Alps (Austria), whereas the Gouffre Mirolda (−1626 m) and the Réseau Jean Bernard (−1602 m) are situated in the French Alps (www.caverbob.com). Shafts may have extreme vertical dimensions, but their diameters are relatively small. The deepest vertical drop with 603 m pitch is situated in Vrtiglavica Cave (Kanin Massif, Slovenia). Shafts are generally created by water flowing from the surface, thus they are called *invasion shafts* (Ford 1983b). Horizontal or subhorizontal passages commonly have elliptical or tubular cross-sections and they are considered epiphreatic or phreatic in origin, i.e. they are formed at or below the karst water table (Ford 1983b; Plan et al. 2009; Bočić et al.

Fig. 5.1 Characteristic shaft in Gortani Cave (Canin Plateau, Italy, photo by Egri)

Fig. 5.2 Phreatic passage in Gortani Cave (Canin Plateau, Italy). Note that the profile is influenced by layer dip and by a vertical fissure as well (photo by Egri)

2012, Fig. 5.2). While horizontal cave passages usually do not show significant vertical amplitudes in the Alps (Audra et al. 2007), there are remarkable vertical excursions in the geometry of caves in the Canadian Rocky Mountains. These latter features are called phreatic loops (Ford 1983b). Another characteristic of glaciokarst caves is that maze sections are often found either in alpine (White 1979; Plan et al. 2009) or in arctic settings (Skoglund et al. 2010; Farrant and Simms 2011; Lauritzen and Skoglund 2013). A special layout of Norwegian karsts is the so-called *stripe karst* that means a narrow band of karstifiable rocks on the surface that is bordered by non-karstic rocks at both sides. It is a result of metamorphic rock structure typical in the Scandinavian Mountains (Horn 1937; Lauritzen and Skoglund 2013). In stripe karst settings, underground passages are usually constrained by a relatively thin karstifiable bed, commonly marble, thus caves are in fact 2D features, but usually not horizontal. Stripe karsts contain only short phreatic passages, occasionally crossed by vadose branches (Faulkner 2006).

Passage shapes are determined by several factors. First, by the hydrological regime, because the shape of the completely water-filled *phreatic* passages and the shape of the only partly and periodically water-filled *vadose* passages are completely different. In the first case, the cross-section is tubular or rounded, while in the second case, the passages are vertically elongated with a *subsurface canyon-like* morphology. As a result of uplift relative to base level, phreatic passages gradually become vadose, therefore, their shapes are transformed and characteristic keyhole-profile may evolve (Chevalier 1944; Ford and Williams 2007; Bočić et al. 2012). Naturally, these cross-sections are not exclusive to glaciokarsts, however, it is true that the hydrologic regime, i.e. the phreatic or vadose formation of passages quite frequently changed in the geological history of glaciokarsts due to glacial cycles. Nonetheless, changes in karst water level can be caused by other factors as well, namely by tectonic uplift or subsidence, or by other climatic changes, and by sea level oscillations. Passage cross-sections are also influenced by general geologic factors, such as bedding planes (see e.g. Ford 1983b), tectonic fissures, joints, fault planes (see e.g. Bodenhamer 2007; Farrant and Simms 2011) or weakness zones due to metamorphosis (see e.g. Lauritzen 1984) similarly to other, non-glaciokarst caves.

A characteristic impact of glacial erosion is the *truncation* of cave segments near the surface. Unroofed caves (called *Ruinenhöhlen* by Lechner in Audra et al. 2002) are transitional forms between surface and subsurface features. They are very common phenomena in the Alps (Audra et al. 2002; Plan et al. 2009), but also in other places like in the Aladağlar Mountains (Turkey), where Klimchouk et al. (2006) discriminated several subtypes of unroofed caves. There are shafts, whose upper parts were completely destroyed by ice including a certain thickness of bedrock that he calls *decapitated shafts* (Fig. 5.3). In some other cases, only the epikarst segment is cleared away, these forms are *epikarst shafts*. Finally, there are *unwalled shafts* (Fig. 5.4), whose wall was partly destroyed by valley entrenchment, they are geologically short-lived phenomena. It is observed that fluvial incision opens rather the (sub)horizontal passages, whereas vertical shafts are principally opened by glacial erosion (Klimchouk et al. 2006).

Due to *glacial erosion*, cave entrances often get to anomalistic position, i.e. the entrance is found hanging high at the valley side (Fig. 5.5), without either spring or sink functionality, if only the present topographic and hydrologic situation is taken into consideration (Oxaal 1914; Ford 1983b; Lauritzen 1984, 1986). Cave entrances can be partially destroyed by direct glacial scouring, but frost shattering is also an active agent. It worked efficiently in the glacial periods, but it is still active in periglacial conditions (Ford 1983b; Murphy et al. 2008, 2015). If destructive effects are considered at a larger scale, then it must be taken into account that valley entrenchment may demolish certain preglacial passages, others may get to different hydrologic positions, and in general, the karst aquifer may become disordered, and formerly contiguous caves may turn to be dissected (Ford 1983a; Ford and Williams 2007).

Among the small-scale dissolutional cave features, *paragenetic forms* are mentioned first (Skoglund and Lauritzen 2011; Farrant and Smart 2011; Bočić et al. 2012). They are created when a passage is filled with sediments, and water finds a way between the sediment and the cave wall, and so dissolves the limestone profile. In many cases, the ceiling is the target of the dissolution. Paragenesis is usually possible under phreatic conditions. Paragenetic features include ceiling half tubes (Fig. 5.6), ceiling meanders and bypasses (Farrant and Smart 2011; Plan et al. 2009). It is important to note that paragenetic features are not restricted to

Fig. 5.3 Shaft entrance at a tectonic fissure and the glacially eroded surface (Tennengebirge, Austria, photo by Egri)

glaciokarst caves, they are present in different climates as well, but they are typical in glaciokarst areas, because the filling of passages by sediments is a frequent event during glacial–interglacial transitions (Farrant and Smart 2011). Other common small scale cave features are the *scallops* (Fig. 5.7), which are created by dissolution due to flowing water in cave passages (Ford 1983b). Again, they are not confined to glaciokarst caves, but they frequently occur in them, and in the lack of other formations (e.g. speleothems), they are often the dominant small-scale features of glaciokarst caves. In order to highlight their significance, it is noted that former flow velocities and directions can be calculated from scallops morphology, therefore former hydrologic conditions can be reconstructed. Notably, the hydraulic gradient can be calculated based on the roughness of scallops (Lauritzen 1986). By the examination of several Norwegian caves, Lauritzen and Skoglund (2013) concluded that relatively low flow velocities created the scallops of the studied caves, and in certain cases, they found that flow directions were just the opposite of what would seem logical based on the present-day topography-related directions. In the area of Yorkshire Dales, Murphy et al. (2015) observed particular *scoops*, which could be also flow generated features. However, as these forms are larger than normal scallops, their development can be explained by faster flow, namely by the water flowing between the glacier ice and the rock wall.

Fig. 5.4 Unwalled shaft at the Canin Plateau (Italy, photo by Telbisz)

In permafrost areas, most of the subsurface water is in a frozen state that means the interruption of karstification. Notwithstanding, there exist some karst terrains in permafrost areas, too. However, their development took place in different climatic conditions (Ford and Williams 2007). Among permafrost terrains, Svalbard has been studied by Lauritzen (2006), who described some caverns as well. These caverns are small and fully plugged with ice or ice-cemented debris in some meters from their entrances, but even these small segments contain some scallops and paragenetic features.

Fig. 5.5 Entrance to Eisriesenwelt Ice Cave (Austria) situated in a hanging position (photo by Egri)

Fig. 5.6 Ceiling half tube in Gortani Cave (Canin Plateau, Italy, photo by Egri)

Fig. 5.7 Phreatic passage with scallops (Lamprechtsofen, Austria, photo by Egri)

5.2.2 Characteristic Glaciokarst Cave Sediments and Depositions

The most peculiar sediment deposited in glaciokarst caves is *ice* (Fig. 5.8). However, the formation of ice is not restricted to glaciokarst caves at all. There are a number of ice caves in temperate climate terrains as well, such as the Dobšinská Ice Cave in the Carpathians (Slovakia) or the Scarisoara Ice Cave in the Apuseni Mountains (Romania), which are found in medium mountains in continental climate. Obviously, the best-known ice caves (e.g. the Eisriesenwelt or the Dachstein Ice Cave, both in the eastern Alps, Austria) are found in glaciokarst terrains, since these are the environments where the present climate is cold enough for the accumulation of ice in the caves. On the contrary, the previously mentioned ice caves in temperate climate medium mountains could be formed only due to specific topographic settings. Turning to actually glacierized karsts, the presence of ice can be even more substantial. The expedition into Castleguard Cave led by Derek Ford explored the cave starting from the spring mouth by going upwards, and when they reached the upstream end, they were the first to discover an ice plug in a cave that completely filled the passage and that was created by the penetration of glacier ice into the cave (Ford 1983b). Moreover, in the Alps and other high karst mountains, caves, especially shafts, which are partly open to the surface are often permanently filled with snow and ice. Depth of such an ice or snow plug may reach several hundred meters. Smart (2004) mentions, for instance that cavers descended down to 200 m deep in a shaft between the ice plug and the rock wall in the Julian Alps

Fig. 5.8 Ice formation in Eiskogelhöhle Cave (Tennengebirge, Austria), which is the highest show cave in Europe (photo by Egri)

(Slovenia), and Plan et al. (2009) described a 250 m deep ice plug from Totes Gebirge (Austria). The ice plugs may be permanent or periodic, occasionally blocking the infiltration, but when the ice is partly decayed, water may flow down, and even paragenetic features may be created this way (see Murphy et al. 2015).

Coarse debris accumulations are commonly found in glaciokarst caves, they are typical of the vadose sections (Audra et al. 2002; Bočić et al. 2012). Source of the coarse debris may be directly the frost-shattered rock, but flowing water may also bring glacial debris and morainic material into the cave (Audra et al. 2002). This latter option requires streams with enough discharge and energy. The sediments detached from glacial debris, transported and reworked by flowing water are called fluvioglacial deposits (Farrant et al. 2014; Bočić et al. 2012). Fluvioglacial deposits are usually rounded or sub-rounded, and the clast material depends on the source, which may be karstic in case of limestone dominated mountains or non-karstic if there are other bedrocks within the glacier domain. Size of the clasts ranges from gravels to boulders and fluvioglacial sediments are usually poorly sorted (Plan et al. 2009; Bočić et al. 2012). These deposits may be porous, but occasionally they may completely fill shafts and less frequently horizontal passages as well (Plan et al. 2009; Bočić et al. 2012). If the bottom of the ice plug melts in a shaft, then the bottom debris gradually becomes cemented, and so-called *cemented diamict* is created (Klimchouk et al. 2004).

Beside the accumulation of coarse debris, *fine-grained sediments* are also frequently deposited in glaciokarst caves (Fig. 5.9). They are called *varved carbonate sediments*, i.e. they are finely laminated and contain carbonate material (Maire 1990; Audra et al. 2002; Plan et al. 2009). The carbonate content often consists of fine calcite flakes, and the deposits occasionally contain high amount of quartz grains, too (Audra et al. 2002, 2004). The typical grain size is silt. Commonly, there are alternating light and dark laminae according to the high seasonal changes of discharge (Audra et al. 2002). The thickness of fine-grained sediments often reaches several meters, and they may effectively clog cave passages. Given the grain size of that material, it is obvious that they are deposited from suspension (Riviere 1977 in: Audra et al. 2002). Typically, there are black-floodings in the caves that causes the deposition of these varved sediments (Audra et al. 2002; Plan et al. 2009; Farrant and Smart 2011; Bočić et al. 2012).

Both the fine and coarse-grained sediments may result in the *clogging* of certain cave passages either near the entrance or deeper within the cave. The deeper parts are mainly plugged by fine-grained sediments, whereas the sections near the entrance are typically blocked by coarse sediments. The entrance of a cave can be directly dammed by glacial till as well (Waltham et al. 1997; Skoglund et al. 2010; Bočić et al. 2012). Sedimentary plugs may be flushed by extreme high discharges, which are characteristic to the end of glacial phases due to the sudden increase of meltwaters or they may occur during interglacial periods in general.

Fig. 5.9 Laminated, fine-grained sediments in Gortani Cave (Canin Plateau, Italy, photo by Egri)

Commonly, the best known and most peculiar cave sediments are *speleothems*. They are relatively rare phenomena in glaciokarst caves with respect to other climatic environments, but they do exist. According to classical theories of speleothem growth, they can form only in caves below soil-covered terrains (Ford and Williams 2007). Thus, even if they occur in some glaciokarst caves, their formation is linked to pre-, inter-, or postglacial warm periods (Audra et al. 2002). Given this general view, it was surprising new information in the 1970s that in the Castleguard Cave, below an actually glacier-covered terrain, only some 150 m from the ice plug at the end of the cave, small speleothems were found by Ford and his companions, and a bit farther away from the ice plug, even larger speleothems were observed, which showed traces of active growth (Ford 1983b). The mechanism of speleothem growth is presented in the following subchapter. The presence of speleothems may provide information about the activity erosion processes as well. In several caves of the Alps, speleothems are well preserved in seemingly active passages. However, the presence of apparently intact speleothems proves that both the mechanical erosion and the chemical aggressivity of waters is at a low level in the given passages since the formation of the speleothems (Audra 2004).

Cryogenic cave calcites (CCC) are rare phenomena, created when the carbonate content of the water precipitates during freezing (Clark and Lauriol 1992; Žák et al. 2004). They accumulate at the bottom of caves, and they consist of relatively small size (1–10 mm), and varied shapes of calcite crystals. The amount of cryogenic carbonates is negligible with respect to other deposits (Spötl and Mangini 2007a, b), but they can be used as climate markers. They are formed either in glacial or in periglacial environments, when liquid water is occasionally present in the karst system but the temperature is usually below zero. They bear useful information for climate chronology because they can be numerically dated (Žák et al. 2009, 2012).

5.3 Subsurface Processes in Glaciokarst Terrains

5.3.1 Age and Formation of Speleothems

As it was already mentioned, speleothem growth was previously linked to warmer (interglacial, preglacial, postglacial or occasionally to interstadial) periods, when the surface was covered with soil (Ford 1976a, b; Harmon et al. 1977; Atkinson et al. 1978). In some cases, it was even stated that active speleothem growth is possible only below the treeline (Lauritzen and Gascoyne 1980). Thus, it was quite unexpected when Ford (1983b) and Gascoyne et al. (1983) described subglacial speleothem formation. However, this new discovery only modified the earlier views, as it is still true that the main speleothem growth periods can be linked to warmer climates. This fact is also supported by the increasing number of speleothem age data (Atkinson 1983; Gascoyne et al. 1983). Flowstones were observed even on Svalbard, their age proved to be interstadial (Lauritzen 2006). On the other hand, it must be emphasized that hiatuses in speleothem growth are caused not only by cold, permafrost periods, but also by the modification of infiltration paths, or by aridity (Lauritzen and Mylroie 2000). The potential of subglacial speleothem formation in the Alps was studied by Spötl and Mangini (2007a, b), who concluded that while the temperature is above freezing point in the cave and the passage is not completely filled with meltwater, speleothem growth is possible. Notably, the MIS 3 was unambiguously colder than the Holocene, however, the rate of speleothem growth was similar during interstadials and stadials. According to his analysis, speleothem growth begins as soon as the glacier above the cave becomes warm-based and promptly interrupts whenever the cave temperature decreases below freezing point. Further on, he observed that there are no unequivocal macroscopic differences in stalagmites and flowstones formed during the warmer Holocene or during the cooler MIS 3 and MIS 7.

Given the above observations, scientists began to seek for chemical explanations for the formation of subglacial speleothems. Dreybrodt (1982) explained calcite precipitation by the temperature effect. According to his theory, water infiltrating from the surface at a temperature near zero, is capable to dissolve a certain amount of $CaCO_3$. When the solution enters the cave and water temperature increases by some °C, the solution becomes supersaturated with respect to $CaCO_3$ and precipitation begins within the cave. He stated that this mechanism can explain the generation of speleothems below glacier-covered terrains and also below surfaces, which are bare due to other reasons. Atkinson (1983) examined Dreybrodt's theory and three other mechanisms as well, and he directly measured the geochemistry of karst waters in Castleguard Cave. One of the possible mechanisms was that the partial pressure of CO_2 is increased in the infiltrating water due to some non-biogenic reasons. However, this possibility could be excluded based on the field measurements. The temperature effect described by Dreybrodt (1982) is also insignificant according to Atkinson's measurements, though it is responsible for some percents of the calcite precipitation. The precipitation caused by the evaporation of the infiltrating water is also minimal, again responsible to few percents only. Finally, the essential process behind calcite precipitation in Castleguard Cave is the so-called *common-ion effect* (Atkinson 1983). The infiltrating water reacts with sulfuric acid originating from the oxidation of pyrite that makes the dissolution of carbonate minerals (calcite and dolomite) possible even at low partial pressure of CO_2. If the solution is already saturated with respect to calcite, and further gypsum or dolomite is dissolved, then precipitation of calcite begins. Thus, in case of appropriate rock composition, this mechanism can explain speleothem formation even below glaciated terrains, until the temperature is above 0 °C in the cave, and the passage is not fully inundated. Based on the analysis of alpine speleothems, Spötl and Mangini (2007a, b) observed the same effect. In his Austrian study area, marble was dissolved by sulphide-oxidation, and sulphates originated from gneiss layers situated above the marble. Besides, the mechanism of dissolution-precipitation, he also studied the stable isotope content of speleothems. He observed that speleothems formed by sulphide-oxidation have lower $\delta^{13}O$ values, i.e. they are closer to the isotope ratio of meltwaters. The relatively high $\delta^{13}C$ values of subglacial speleothems are even more remarkable, that suggests their formation without soil. Using stable isotope and age data of speleothems together, it is possible to determine when the glaciers above the cave were warm-based (Spötl 2007). According to the above facts, speleothems sensibly react to temperature changes around 0 °C and to hydrological changes caused by the glacier, thus they are appropriate means to study changes in the mass balance of alpine glaciers (Luetscher et al. 2011).

Speleothem growth data can be also used to infer valley entrenchment rates of the surrounding terrain. The basic principle behind this calculation is that speleothem growth can start only if the given passage is above the local karst water table. Hence, the level of the karst water table is often related to the base level of the valley bottom, thus speleothem age data also provides a minimum age for a given level. This principle was used for Norwegian, Canadian and alpine study areas by Lauritzen and Gascoyne (1980), Ford et al. (1981), and Häuselmann et al. (2008), respectively.

5.3.2 Hydrological Characteristics of Glaciokarsts

In case of remote, rarely populated areas, the exploration of the karst hydrological system is important mainly from a scientific point of view, but in the densely inhabited surroundings of high karst terrains, notably in many parts of the Alps, the hydrologic knowledge of glaciokarsts is of utmost importance from the viewpoint of water management. As an example, Vienna (Austria) gets 60% of its water from alpine karst aquifers (Smart 2004; Plan and Decker 2006).

One of the most remarkable characteristics of alpine type glaciokarst hydrology is the high fluctuation of discharge. The first type of fluctuation is due to seasonality, peak discharges are usually the results of the rainfall combined meltwater periods, i.e. they occur generally in summer (Skoglund and Lauritzen 2010). The second type of fluctuation means daily oscillations according to daily changes of icemelt. Naturally, based on the local hydrologic properties of the aquifer, there may be a certain lag between the highest melting time and the spring discharge. Peak discharges often cause flood conditions, occasionally completely filling certain passages, thus a significant proportion of glaciokarst caves can be safely visited only during winter. Among others, summer floods in Castleguard Cave also caused serious situations in the early period of explorations (Ford 1983b).

Based on hydrologic measurements, Smart (1983) classified waters in glaciokarsts into two classes. The first type includes *supraglacial meltwaters*, which cause high daily fluctuations in discharge. In addition, these waters are able to dissolve a certain amount of limestone, and through large sink points, they can transport coarse debris into the cave. The second includes the *subglacial meltwaters*, which are the results of high pressure below warm-based glaciers. They infiltrate into the karst principally through smaller fissures, and occasionally, they may lead to speleothem formation if conditions of the above-described mechanism are satisfied. Both types of waters are active mostly in summer, but in exceptional cases, a small quantity of liquid water may be generated from the glacier even in winter. Smart (1983) identified three conditions for the active functioning of karst hydrology. First, supraglacial meltwater must reach the bottom of the ice—it requires joints and fissures in the glacier ice. Second, it is important that there is no impervious glacial drift below the glacier that could hinder the penetration of water into the bedrock. Third, the hydraulic gradient must be high enough to make water flowing.

The aforementioned varved deposits can be explained by the extreme high discharges in the summer period (Maire 1990; Audra et al. 2002, 2004). Due to the high amount of meltwater flowing into the passages, the karst water level may increase by several hundreds of meters (Ford 1983b). The fine-grained, partly carbonate sediments are transported by the flood to different parts of the cave, and when the water is drawn back, then the silty material gradually settles out from the suspension. In some parts of the cave, temporary lakes may be formed from the remaining flood water. As floods are typical in glaciokarst settings, these fine-grained sediments can be the markers of glacial effects (Maire 1990).

By geochemical analysis, Ek (1964) and Ford (1971) ascertained that the dissolution capacity of subglacial waters is extremely low with respect to supraglacial meltwaters. Low dissolution capacity was also demonstrated by Skoglund and Skoglund and Lauritzen (2013) in Norwegian karst water samples, where they found total hardness values of 15–36 $CaCO_3$ equivalent mg/l.

5.3.3 Speleogenesis in Glaciokarst Terrains

As for the speleogenesis of glaciated karsts, there existed quite different views simultaneously for a long time. Based on the observation that caves and related karst forms are so abundant in the Alps, Maire (1978) suggested that speleogenesis should have been very active during glacial periods. The basic reason behind this activity could be the high amount of water, which flowed into the karst during seasonal snowmelts contributing to the development of cave passages. Corbel (1957), arguing by the fact that there is a high density of caves in the arctic glaciokarst terrains near the margins of the former continental ice sheet, also hypothesized intense karstification during glacial periods, that he explained by high corrosion rates. His theory was later unambiguously refuted by the measurements of Ek (1964) and Ford (1971). Jennings (1971 in Cooper and Mylroie 2015) supposed that in glaciated karsts, all caves are relatively young, i.e. postglacial, because glaciers necessarily destroyed preglacial (or interglacial) forms. Although glacier erosion is really significant, and in some places more than 100 m thick material was eroded by the continental ice sheet (Mangerud et al. 2011; Braun 1989), but the thickness of erosion is highly dependent on topography, ice thickness, ice movement and thermal conditions at the glacier base, and there are areas, where glacial erosion remained only minimal. Later, measurements refuted Jennings's opinion, too. According to Horn (1935), caves are generally small in Scandinavia, so he concluded that glaciokarst caves are generally young and there are certainly no preglacial caves in Norway. However, the small size of caves can be also explained by the low rates and by the fact that cave development was interrupted several times during the Pleistocene (Lauritzen 1986).

Until the 1970s, karst researchers had only indirect ways to imply the age of glaciokarst caves (Waltham 1974 in Cooper and Mylroie 2015), and a more precise estimation of the age of glaciokarst speleogenesis became available only after quantitative dating methods were developed. Speleothem dating has been already discussed above, but additionally, since the 2000s, the age of other cave sediments became also measurable with the help of cosmogenic nuclides. The cosmogenic ^{10}Be and ^{26}Al nuclides occur in quartz grains, and they are often washed into caves. But as they get to the cave, they are shielded from further cosmogenic radiation, and the amount of ^{10}Be and ^{26}Al starts to decay by radioactivity. As the half-times of ^{10}Be and ^{26}Al are different, the ratio of the two decays can be used to infer the date, when the sediment got into the cave (Häuselmann 2007). Notwithstanding, like speleothem age data, cosmogenic nuclides also provide a minimum age for the cave formation, that may be significantly younger than the true age of the cave. A particular advantage of the cosmogenic nuclide methodology, that even several million years old burial ages can be determined, while the upper limit of the U-series dating method, which is generally applied to speleothems, is only around 750 ka (Häuselmann and Granger 2005).

When scientists started to use quantitative speleothem dating methods, the main goal was to determine the eldest possible formations. These results quickly evidenced that either in arctic glaciokarsts (Harmon et al. 1975, 1977; Ford 1976a, b; Atkinson et al. 1978, 1987; Gascoyne et al. 1981; Gascoyne and Ford 1984; Lauritzen and Gascoyne 1980; Lauritzen 1983, 1984), or in alpine glaciokarsts (Audra and Quinif 1997; Spötl et al. 2002a, b; Spötl and Mangini 2007a, b; Holzkämper et al. 2005; Audra et al. 2007; Häuselmann 2007; Luetscher et al. 2011), some speleothems are older than the last glacial period, i.e. they survived at least one glaciation, which means, of course, only a minimum age for the formation of the caves. Later on, especially due to cosmogenic nuclide studies, de facto Pliocene age of certain caves has been also proved, for instance to Siebenhengste cave in Switzerland (Häuselmann et al. 2008). Thus, at present, the most widely accepted view is that the majority of caves are older than the last glacial, and in many cases, they are even preglacial. It is generally valid for the glaciokarsts in the Alps (Audra et al. 2007, Fig. 5.10), the British Isles (Farrant et al. 2014), Norway (Skoglund and Lauritzen 2010) and North America (Atkinson 1983; Burger 2004) as well.

Naturally, there are exceptions, too, i.e. there are caves, whose postglacial age were reliably demonstrated (Cooper 2014; Lauritzen 1986). The postglacial age cannot be attested purely by speleothem data, since the lack of old speleothems or even the absolute lack of speleothems is not a strict evidence for the young age of a cave. Instead, other morphological markers were used to prove the postglacial age. An important observation is that young fissures are often generated in rocks due to postglacial isostatic rebound. These fissures are sometimes enlarged by karst processes. If the fissure is postglacial by genetics, then the cave formed from it must be

Fig. 5.10 Wide, possibly old (preglacial) passage with ice accumulation in Eiskogelhöhle Cave (Tennengebirge, Austria, photo by Egri)

also postglacial in age. Some examples for such features are described by Cooper (2014) in the Appalachian Mountains, and by Faulkner (2006) in Scotland. Similarly, the walls of alpine type glacier troughs are often weakened after glacier withdrawal and stress-release sheet fractures may be created parallel with the valley side. Such forms are presented by Lauritzen (1986), Skoglund and Lauritzen (2010) in Norway, and by Murphy et al. (2015) in Britain. In some other places, caves are related to subglacially created drumlins. For example, if a drumlin blocks a former cave entrance, then a new bypass branch may develop. Or runoff water may create new stream sinks along the edge of drumlins, and thus new cave passages may evolve connected to these sink points. As drumlins were formed below the continental ice sheet, drumlin-related cave passages are necessarily postglacial (Cooper and Mylroie 2015).

In principle, the age of cave sediments (speleothems, coarse debris, fine-grained deposits) is not enough to determine the exact age or the formation period of a given cave passage. Speleogenesis can be related to warm (i.e. preglacial, interglacial or Holocene), cold (i.e. subglacial) or transitional (i.e. deglacial or proglacial) periods. Audra et al. (2007) concluded that in the Alps, the main periods of speleogenesis and of speleothem growth were the interglacials. On the contrary, glacials can be characterized principally by destruction and by filling of passages by sediments. Real speleogenesis, i.e. the formation of new passages is of minor importance during glacial periods. New passages and shafts can be formed only where there are holes in the glaciers (Audra 2004), or where the non-karstic cover layers are eroded by the glaciers and karst bedrocks come to light constituting the new surface. This latter possibility is called the *indirect role* of glaciers by Audra (2004). Ford (1983c), after carefully examining glaciokarsts of the Canadian Rocky Mountains, stated that none of the essential parts of Castleguard Karst can be considered truly preglacial, instead, glacial and karstic processes jointly formed the landscape. Burger (2004) identified three phases of glaciokarst processes based on his research in Colorado. First, during glacial periods, calcite deposition and roof breakdown are characteristic, second, in periglacial conditions, active stream erosion and filling of passages are typical due to high energy meltwaters transporting loads of sediment into the cave, and third, interglacials are characterized by roof collapse, calcite precipitation and sediment deposition. Similar phases have been outlined by Murphy et al. (2008) using Gaping Gill Cave (Yorkshire, UK) as an example. Block breakdowns mostly occur during interglacials, but discharges are not high enough to dissolve and transport the large blocks. Glacial periods are characterized by intense frost shattering, while the transportation and the distribution of the shattered material within the cave is related to deglaciation phases.

The idea of speleogenesis related to high discharges caused by meltwaters during deglaciations is not new, it was already suggested by the Norwegian Oxaal in 1914. When glaciers retreated, meltwaters created huge lakes, which were occasionally dammed by ice. The waters of these lakes had higher dissolution capacity than waters of subglacial origin, thus, where the water of these ice-dammed lakes got into the karst, speleogenesis could be significant due to the high discharge and the relatively high dissolution capacity. This process was influenced by the remarkable fluctuations of the water level of these lakes. Faulkner (2006) and Murphy et al. (2015) stated that this factor could play an important role in the development of several caves in the British Isles and Norway. Faulkner (2006) and Murphy et al. (2015) also suggested that extreme discharge events, such as jökulhlaups could also contribute to the formation of caves and gorges.

The idea of *subglacial speleogenesis* emerged also early in the history of glaciokarst research (Horn 1935, 1947). The observation of Werenskiold (1953) that there is no permafrost below glaciers wider than 400 m provided a further argument in favour of subglacial cave development. However, the demonstration that subglacial waters are usually saturated, and at very low concentrations became a strong argument against the hypothesis of subglacial speleogenesis (Ek 1964; Ford 1971). Lauritzen and Skoglund (2013) and Skoglund et al. (2010) calculated by computer simulation that subglacial wall retreat is cca. 50 times slower than wall retreat during interglacials. Moreover, the breakthrough time of cave inception is 26 times longer in glacials than in interglacials. Nevertheless, cave development is not completely interrupted during glacials, only during permafrost periods. In the Norwegian karst areas studied by Lauritzen and Skoglund, there were large enough glaciers during a significant part of the Quaternary, when permafrost was not present below the glaciers, therefore, cave development was possible. In case of non-karstic rocks, the erosion of subglacial waters creates *Nye-channels*, which mean relatively small streambeds incised into the bedrock. In case of karstic bedrock, these Nye-channels are specifically transformed and may be occasionally created as small, near surface cave passages (Lauritzen 1986). However, the dissolution caused by supraglacial waters is much more significant. These waters enter the karst at ice margins, so Lauritzen and Skoglund (2013) called their theory *ice-contact speleogenesis*. Supraglacial waters reach the karst bedrock either directly from the ice surface, or from lower parts of the glacier ice due to crevasses. The different types and sizes of crevasses in the glacier form a glacial aquifer, which is, in turn, joint to the karst aquifer formed within the bedrock. Whenever the subsurface flow is directed towards the rock, speleogenesis becomes more intensive, and due to diffuse recharge conditions, the cave evolves to a maze type (cf. Palmer 2003), which is a frequent phenomenon in Norwegian karsts. This process could be manifested by computer simulation as well (Skoglund et al. 2010). Scallops also demonstrate the existence of a joint aquifer between the ice and the rock mass. Low flow velocities calculated based on the scallops imply low hydraulic gradients, which is feasible only if the glacial and karstic aquifers were connected. Moreover, flow directions opposed to the present-day topographic gradients can be most easily explained if the glacier surface is also taken into account (Lauritzen 1986). Depending on the ice thickness of the glacier, even the same passages could have opposite flow directions in different glacial periods, that sounds paradoxical at first. The key is the ice thickness, which determines the hydraulic gradient. This occasionally alternating flow direction is called *see-saw effect* by Farrant and Simms (2011) and by Farrant et al. (2014).

The style of cave development is largely determined by the karst water level, as it determines whether phreatic or vadose processes form a certain passage. On the other hand, hydraulic gradient is also highly important. Based on these factors, alpine type glaciokarsts can be grouped into two types (Ford and Williams 2007). The first is the *Canadian type alpine glaciokarst*, where glaciers are present both in the higher zones and in the lower zones of the mountains, in the valleys, occasionally spreading to the piedmont area as well. This arrangement results that not only the water recharge zone but also the spring zone of the karst is affected. The valleys filled with glaciers and partly by till block the outflowing springs and usually cause black-flooding in the karst aquifer. Moreover, melting of glaciers provides an extra water recharge into the karst system. Consequently, a remarkable karst water level increase can be generally attributed to glacial periods, that significantly influences speleogenesis as well (Ford 1983b; Audra 2004). Further on, it is noted that if the karst aquifer is dammed at the output side, it may also cause the development of maze sections (Palmer 2003; Lauritzen and Skoglund 2013). The Canadian type is characteristic in the Alps, the Scandinavian Mountains, and in the Canadian Rocky Mountains from where it got its name (Ford and Williams 2007). The second group is the *Pyrenean type alpine glaciokarst*, where glaciers are restricted to the higher zones, while the lower parts of the valleys remain unglaciated. In this type, the outflow is not hindered, thus glacial periods do not cause significant water level increase. This group includes obviously the Pyrenees, the southern parts of the Alps, the Taurus, the Caucasus and the US part of the Rocky Mountains (Ford and Williams 2007). Nevertheless, Audra et al. (2007) remarked that the characteristic cave horizons present in many large alpine systems cannot be interpreted as parts of a former, contiguous regional karst water level because water level fluctuations are influenced more locally by individual valley glaciers.

Glacial erosion is one of the most effective forms of erosion, so alpine valley entrenchments were remarkable during glacial periods, therefore, erosion base levels and, consequently, karst springs were significantly lowered after deglaciation (Ford and Williams 2007; Skoglund and Lauritzen 2010). Speleogenesis usually follows these relatively abrupt base level drops by a certain time lag.

Finally, it is necessary to mention another significant factor influencing base level, that is not related at all to glacial processes. However, it affected many glaciokarst terrains at the southern side of the Alps. It is observed in many places that caves were adapted to a much lower karst water table and consequently to a much lower erosion base level. The reason for this is the so-called *Messinian Salinity Crisis*, during which the Mediterranean Sea nearly completely desiccated between 6 and 5.3 Ma ago. This event lowered base levels in some places by more than 1000 m, so it must be taken into consideration when cave horizons are interpreted, but should not be confused with “normally” low interglacial base levels (Audra 2004).

5.4 Conclusions

Glaciokarst terrains are often rich in caves. In many cases, the exploration of these caves was a real challenge to cavers and posed several scientific questions to karst researchers. Among glaciokarst caves, there are many compounds, very long systems, but passage volumes are generally not so large. On the other hand, vertical dimensions are remarkable, and most of the deepest caves on Earth are found in glaciokarst terrains (in the Caucasus, the Alps, and the Pyrenees).

Besides, the thorough morphological description of caves, mainly the cave sediments (speleothems, coarse debris, fine-grained deposits) helped scientists to decipher the age of speleogenesis, that, in turn, provided the key to understand the mechanisms of cave development. Due to modern, quantitative dating methodology, namely U-series dating and cosmogenic nuclide techniques, it is now unequivocal that caves of the different glaciokarst terrains have a wide range of age, there are postglacial caves as well, but most of them are significantly older.

Field observations, geochemical analysis, compound models and computer simulations all contributed to demonstrate that speleothem growth and even speleogenesis is possible in subglacial environments if liquid water exists in the karst system. However, it allows only limited growth and enlargement, whereas speleothems and cave passages increase dominantly during warmer (preglacial, interglacial or Holocene) periods. As for the relative significance of glacial or deglacial periods, opinions are divided so far, and it is feasible that there is not a global answer, but situations are locally different.

In general, it is stated that glaciokarst terrains were repeatedly affected by Quaternary climate fluctuations, and as the majority of caves are older than the last glacial period, their development is polygenetic in the meaning of Ford and Williams (2007). During cave evolution, the effects of several phases have been stacked together (Skoglund and Lauritzen 2010), but these phases were not equal neither in duration nor in their speleogenetic significance. Taken into consideration permafrost periods as well, it is clear that cave development was intermittent in most glaciokarst areas (Skoglund et al. 2010). The image is even more complicated in tectonically active areas, where the effects of climate change and tectonic uplift, notably karst water level changes, are intermingled (Bočić et al. 2012). This is why Häuselmann et al. (2008) emphasized that the interpretation of paleoclimate or paleogeography from cave information is not straightforward!

References

Atkinson TC, Harmon RS, Smart PL, Waltham AC (1978) Palaoclimatic and geomorphic implications of ^{230}Th/^{234}U dates on speleothems from Britain. Nature 272:24–28

Atkinson TC, Lawson TJ, Smart PL, Harmon RS, Hess JW (1987) New data on speleothem deposition and palaeoclimate in Britain over the last forty thousand years. J Quat Sci 1:67–72

Atkinson TC (1983) Growth mechanisms of speleothems in Castleguard Cave, Columbia Icefields, Alberta Canada. Arctic Alp Res 15(4):523–536

Audra P, Bini A, Gabrovšek F, Häuselmann P, Hobléa F, Jeannin PY, Kunaver J, Monbaron M, Šušteršič F, Tognini P, Trimmel H, Wildberger A (2007) Cave and karst evolution in the Alps and their relation to paleoclimate and paleotopography. Acta Carsologica 36(1):53–68

Audra P, Quinif Y (1997) Une cavité de haute-montagne originale: la grotte Téophile (Alpe d'Huez, France) Rôle des paléoclimats pléistocènes dans la spéléogenèse. Spéléochronos 8:23–32

Audra P (1994) Alpine karst speleogenesis: case studies from France (Vercors, Chartreuse, Ile de Crémieu) and Austria (Tennengebirge). Cave Karst Sci 21(3):75–80

Audra P (2004) An overview of the current research carried out in the French Western Alps karsts. Acta Carsologica 33(1):25–44

Audra P, Quinif Y, Rochette P (2002) The genesis of the Tennengebirge karst and caves (Salzburg, Austria). J Cave Karst Stud 64(3):153–164

Audra P, Mocochain L, Camus H, Gilli É, Clauzon G, Bigot J (2004) The effect of the Messinian Deep Stage on karst development around the Mediterranean Sea. Examples from Southern France. Geodinamica Acta 17(6):27–38

Bočić N, Faivre S, Kovačić M, Horvatinčić N (2012) Cave development under the influence of Pleistocene glaciation in the Dinarides—an example from Štirovača Ice Cave (Velebit Mt., Croatia). Zeitschrift für Geomorphologie 56(4):409–433

Bodenhamer HG (2007) Preglacial development of caves at structural duplexes on the Lewis Thrust, Glacier National Park, Montana. J Cave and Karst Stud 69(3):326–341

Braun DD (1989) Glacial and periglacial erosion of the Appalachians. Geomorphology 2(1–3):233–256

Brook GA, Ford DC (1980) Hydrology of the Nahanni Karst, northern Canada, and the importance of extreme summer storms. J Hydrol 46(1–2):103–121

Burger PA (2004) Glacially-influenced sediment cycles in the Lime creek karst, Eagle County, Colorado. In Studies of cave sediments, Springer, pp 107–122

Chevalier P (1944) Distinctions morphologiques entre deux types d'érosion souterraine. Revue de géographie alpine 32(3):475–486

Clark ID, Lauriol B (1992) Kinetic enrichment of stable isotopes in cryogenic calcites. Chem Geol Isot. Geosci Sect 102:217–228

Cooper M (2014) Verification of post-glacial speleogenesis and the origins of epigene maze caves in New York. MSc dissertation, Mississippi State University

Cooper MP, Mylroie JE (2015) Glaciation and Speleogenesis. Springer

Corbel J (1957) Les karsts du Nord-Ouest de l'Europe et de quelques régions de comparaison: étude sur le rôle du climat dans l'érosion des calcaires. Revue de Géographie de Lyon 12

Dreybrodt W (1982) A possible mechanism for growth of calcite speleothems without participation of biogenic carbon dioxide. Earth Planet Sci Lett 58(2):293–299

Ek C (1964) Note sur les eaux de fonte des glaciers de la Haute Maurienne (Savoie, France): leur action sur les carbonates. Revue belge de Géographie 88(1–2):127–156

Farrant AR, Simms MJ (2011) Ogof Draenen: speleogenesis of a hydrological see-saw from the karst of South Wales. Cave Karst Sci 38(1):31–52

Farrant AR, Smart PL (2011) Role of sediment in speleogenesis; sedimentation and paragenesis. Geomorphology 134(1):79–93

Farrant AR, Smith CJ, Noble SR, Simms MJ, Richards DA (2014) Speleogenetic evidence from Ogof Draenen for a pre-Devensian glaciation in the Brecon Beacons, South Wales UK. J Quat Sci 29(8):815–826

Faulkner T (2006) The impact of the deglaciation of central Scandinavia on karst caves and the implications for Craven's limestone landscape. In: Re-thinking Craven's limestone landscape, p 4

Ford DC (1971) Alpine Karst in the Mt. Castleguard-Columbia icefield area, Canadian rocky mountains. Arctic Alp Res 15(4):239–252

Ford DC (1976a) Evidences of multiple glaciation in South Nahanni National Park, Mackenzie Mountains, Northwest Territories. Can J Earth Sci 13(10):1433–1445

Ford DC (1976b) Evidence of multiple glaciation in South Nahanni National Park, Mackenzie Mountains, Northwest Territories. Can J Earth Sci 13:1433–1445

Ford DC (1983a) Effects of glaciations upon karst aquifers in Canada. J Hydrol 61(1–3):149–158

Ford DC (1983b) The physiography and speleogenesis of castleguard cave, Columbia icefields, Alberta Canada. Arct Alp Res 15(4):437–450

Ford DC (1983c) Alpine karst systems at crowsnest pass, Alberta-British Columbia Canada. J Hydrol 61(1):187–192

Ford DC, Williams P (2007) Karst hydrogeology and geomorphology. John Wiley and Sons Ltd, West Sussex, England

Ford DC, Schwarcz HP, Drake JJ, Gascoyne M, Harmon RS, Latham AG (1981) On the age of the extant relief in the southern Rocky Mountains of Canada. Arct Alp Res 13(1):1–10

Gascoyne M, Ford DC (1984) Uranium series dating of speleothems. II: results from the Yorkshire Dales and implications for cave development and quaternary climates. Cave Sci 11(2):65–85

Gascoyne M, Currant AP, Lord TC (1981) Ipswichian fauna of Victoria Cave and the marine palaeoclimatic record. Nature 294(5842):652–654

Gascoyne M, Latham AG, Harmon RS, Ford DC (1983) The antiquity of Castleguard Cave, Columbia Icefields, Alberta Canada. Arc Alp Res 15(4):463–470

Harmon RS, Ford DC, Schwarz HP (1977) Interglacial chronology of the Rocky and MacKenzie Mountains based upon $^{230}Th/^{234}U$ dating of calcite speleothems. Can J Earth Sci 14:2543–2552

Harmon RS, Thompson P, Schwarcz HP, Ford DC (1975) Uranium-series dating of speleothems. Nat Speleol Soc Bull 37:21–33

Häuselmann P, Granger DE (2005) Dating of caves by cosmogenic nucleides: method, possibilities, and the Siebenhengste example (Switzerland). Acta Carsologica 34(1):43–50
Häuselmann P, Lauritzen SE, Jeannin PY, Monbaron M (2008) Glacier advances during the last 400 ka as evidenced in St. Beatus Caves (BE, Switzerland). Quat Int 189(1):173–189
Häuselmann P (2007) How to date nothing with cosmogenic nuclides. Acta Carsologica 36(1): 93–100
Holzkämper S, Spötl C, Mangini A (2005) High-precision constraints on timing of Alpine warm periods during the middle to late Pleistocene using speleothem growth periods. Earth Planet Sci Lett 236(3):751–764
Horn G (1935) Űber die Bildung von Karsthöhlen unter einem Gletcher. Nor Geogr Tidsskr 5:494–498
Horn G (1937) Über einige Karsthöhlen in Norwegen. Mitteilungen für Höhlen und Karstforschung 1–15
Horn G (1947) Karsthuler i Nordland. Norg Geol Unders 165:1–177
Klimchouk A, Bayari S, Nazik L, Törk K (2006) Glacial destruction of cave systems in high mountains, with a special reference to the Aladaglar massif, Central Taurus Turkey. Acta Carsologica 35(2):111–121
Klimchouk A, Nazik L, Bayari S, Tork K, Kasjan Y (2004) Kuzgun Cave and its Context: the first super-deep cave in the Aladaglar Massif, Turkey. www.researchgate.net
Lauritzen SE (1983) Arctic and alpine karst symposium. Program and field guide. Dept. of Chemistry, University of Oslo. 89 p
Lauritzen SE (1984) Evidence of subglacial karstification in Glomdal, Svartisen, Norway. Nor Geogr Tidsskr 38:169–170
Lauritzen SE (1986) Kvithola at Fauske; Northern Norway: an example of ice- contact speleogenesis. Nor Geol Tidsskr 66:153–161
Lauritzen SE, Gascoyne M (1980) The first radiometric dating of Norwegian stalagmites–Evidence of pre-Weichselian karst caves. Nor Geogr Tidsskr 34:77–82
Lauritzen SE (2006) Caves and speleogenesis at Blomstrandsøya, Kongsfjord W. Spitsbergen. Int J Speleol 35(1):37–58
Lauritzen SE, Mylroie JE (2000) Results of a speleothem U/Th dating reconnaissance from the Helderberg plateau New York. J Cave Karst Stud 62(1):20–26
Lauritzen SE, Skoglund RØ (2013) Glacier ice-contact speleogenesis. Treatise Geomorphol 6:363–396
Luetscher M, Hoffmann DL, Frisia S, Spötl C (2011) Holocene glacier history from alpine speleothems, Milchbach cave Switzerland. Earth Planet Sci Lett 302(1):95–106
Maire R (1978) Les karsts sous-glaciaires et leurs relations avec le karst profond. Revue de géographie alpine 66(2):139–148
Maire R (1990) La haute montagne calcaire: karsts, cavités, remplissages, paléoclimats, Quaternaire. Karstologia-Mémoires 3:1–731
Mangerud J, Gyllencreutz R, Lohne Ö, Svendsen JI (2011) Glacial history of Norway. In: Ehlers J, Gibbard PL, Hughes PD (eds) Quaternary glaciations-extent and chronology: a closer look. Elsevier
Murphy PJ, Faulkner TL, Lord TC, Thorp JA (2015) The caves of Giggleswick Scar-examples of deglacial speleogenesis? Cave Karst Science 42(1):42–53
Murphy P, Westerman AR, Clark R, Booth A, Parr A (2008) Enhancing understanding of breakdown and collapse in the Yorkshire Dales using ground penetrating radar on cave sediments. Eng Geol 99(3):160–168
Oxaal L (1914) Kalkstenshuler i Ranen. Norg Geol Unders 69:1–47
Palmer AN (2003) Speleogenesis in carbonate rocks. Speleogenesis Evol Karst Aquifers 1(1):2–11
Palmer AN (1991) Origin and morphology of limestone caves. Geol Soc Am Bull 103(1):1–21
Plan L, Decker K (2006) Quantitative karst morphology of the Hochschwab plateau, Eastern Alps, Austria. Zeitschrift Für Geomorphologie Supplementband 147:29–54

Plan L, Filipponi M, Behm M, Seebacher R, Jeutter P (2009) Constraints on alpine speleogenesis from cave morphology—a case study from the eastern Totes Gebirge (Northern Calcareous Alps, Austria). Geomorphology 106(1):118–129

Salvigsen O, Elgersma A (1985) Large-scale karst features and open taliks at Vardeborgsletta, outer Isfjorden. Svalbard, Polar Res 3(2):145–153

Skoglund RØ, Lauritzen SE (2010) Morphology and speleogenesis of Okshola (Fauske, northern Norway): example of a multi-stage network cave in a glacial landscape. Norw J Geol 90:123–137

Skoglund RØ, Lauritzen SE, Gabrovšek F (2010) The impact of glacier ice-contact and subglacial hydrochemistry on evolution of maze caves: a modelling approach. J Hydrol 388(1):157–172

Skoglund RØ, Lauritzen SE (2011) Subglacial maze origin in low-dip marble stripe karst: examples from Norway. J Cave Karst Stud 73(1):31–43

Skoglund RØ, Lauritzen SE (2013) Characterisation of a post-glacial invasion aquifer: the Grønli-Seter karst system, northern Norway. Norwegian J Geol 93(1):61–73

Smart CC (2004) Glacierized and glaciated karst. Gunn J (edt) Encyclopedia of caves and karst science New York. Fitzroy Dearborn, NY, pp 804–809

Smart CC (1983) The hydrology of the castleguard karst, columbia icefields, Alberta Canada. Arctic Alp Res 15(4):471–486

Spötl C, Mangini A (2007a) Speleothems and paleoglaciers. Earth Planet Sci Lett 254(3):323–331

Spötl C, Mangini A (2007b) Speleothems and paleoglaciers. Earth Planet Sci Lett 254(3):323–331

Spötl C, Mangini A, Frank N, Eichstädter R, Burns SJ (2002a) Start of the last interglacial period at 135 ka: evidence from a high Alpine speleothem. Geology 30(9):815–818

Spötl C, Unterwurzacher M, Mangini A, Longstaffe FJ (2002b) Carbonate speleothems in the dry, inneralpine Vinschgau valley, northernmost Italy: witnesses of changes in climate and hydrology since the last glacial maximum. J Sediment Res 72(6):793–808

Szabó L (2008) Barlangfejlődés a Canin-fennsík mélyén. Karsztfejlődés 13:247–267

Szabó L (2009) Cave area of the Canin-Plateau—a naturally geodiverse land in the middle of Europe. Acta Climatologica et Chorologica 42–43:143–150

Waltham AC, Simms MJ, Farrant AR, Goldie HS (1997) Karst and caves of Great Britain. Geol Conserv Rev 12, 358 p

Werenskiold W (1953) The extent of frozen ground under the sea bottom and glacier beds. J Glaciol 2:197–200

White WB (1979) Karst landforms in the Wasatch and Uinta Mountains, Utah. NSS Bulletin 41:80–88

Žák K, Hercman H, Orvošová M, Jačková I (2009) Cryogenic cave carbonates from the Cold Wind Cave, Nízke Tatry Mountains, Slovakia: Extending the age range of cryogenic cave carbonate formation to the Saalian. Int J Speleol 38(2):139–152

Žák K, Richter DK, Filippi M, Živor R, Deininger M, Mangini A, Scholz D (2012) Cryogenic cave carbonate–a new tool for estimation of the Last Glacial permafrost depth of the Central Europe. Clim Past Discuss 8(3):2145–2185

Žák K, Urban J, Cílek V, Hercman H (2004) Cryogenic cave calcite from several Central European caves: age, carbon and oxygen isotopes and a genetic model. Chem Geol 206(1):119–136