

GUIDEBOOK

Cenozoic freshwater carbonates of the Central Carpathians (Slovakia): facies, environments, hydrological control and depositional history

Guide to field trip B7 • 26–28 June 2015

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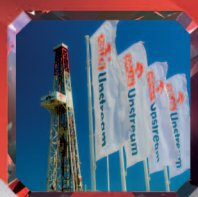
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Cenozoic freshwater carbonates of the Central Carpathians (Slovakia): facies, environments, hydrological control and depositional history

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Route (Fig. 1): From Kraków we follow E77 (S7) south to Rabka, then 47 to Nowy Targ and in town we turn left onto 49. On leaving the town road 49 turns south and crosses the boundary with Slovakia between Jurgów and Podspády. In Slovakia we follow road 67 to **Ždiar Pass (stop B7.1)** with a car park on the right. We follow to Spišská Belá and turn left onto road 77. At Nižné Ružbachy we turn left onto a local road to **Vyšné Ružbachy (stop B7.2)**. From Vyšné Ružbachy we return to Spišská Belá and take road 77 to Kežmarok. There we turn left onto local road 536 to Jánovce and turn east onto road 18. Leaving on the left the mediaeval town of Levoča (with Spišská Kapitula and Spišský hrad parts of the UNESCO World Heritage) we drive 10 km east, leaving aside the entrance on the motorway, to **Sivá brada (stop B7.3)** on the right, at a junction with a side road lined with tall trees. From Sivá brada we drive the side road eastward and then south through the medieval town Spišská Kapitula and descend to Spišské Podhradie. We turn right, continue south, with a travertine ridge on the left, and reach the headquarters of the Euro Kameň company. From the headquarters we follow along the slope of the

north-south trending travertine ridge, pass by a small active quarry to reach inactive **Žehra quarry (stop B7.4)**, at the southern end of the travertine ridge. We depart Žehra quarry and the headquarters of the Euro Kameň company by road 547 south, to Spišské Vlchy. There we turn right to road 536 and continue west to Spišská Nová Ves, where we turn left and proceed south, crossing a forested mountain chain of the Slovak Ore Mountains. We reach road 67 and continue SE towards Rožňava (in Hungarian – Rozsnyó, in German – Rosenau), which is a centre of the historic region called Gemer. The city has an old mining tradition; silver, gold and especially iron ores were exploited there. On the second day, after an overnight stay in Rožňava, the trip departs via the village of Jovice and arrives at the village of Krásnohorská Dlhá Lúka. We reach **Buzgó stream (stop B7.5)** by a 650 m walk along the foot of the Silica Plateau. From the village of Krásnohorská Dlhá Lúka we drive east, along a local road and next turn right onto road 50 and after crossing a pass between Silica Plateau and Horný Vrch Plateau we descend to the Turňa Basin and continue eastward. After 23 km we turn left onto a local

road and pass by the village of Háj (in Hungarian – Áj) to enter a narrow **Háj Valley (stop B7.6)**, with outcrops of inactive Holocene tufa and modern tufa sites. After visiting the Háj Valley we drive back through Rožňava and across the Slovak Ore Mountains to the foot of the Tatras. Then, we enter motorway D1 and continue east to the village of **Bešeňová (stop B7.7)**. After visiting two localities in Bešeňová we drive back to D1 and east to Liptovský Mikuláš, where we turn onto road 584 to reach the outskirts of the village of Demänová, at the foot of the Low Tatras. After an overnight stay in Demänová, we drive south into the Low Tatras 4 km up the Demänová Valley to visit **Demänová Cave System (stop B7.8)**. From there we drive back to Liptovský Mikuláš, then westward by D1 to Bešeňová, where we turn onto a local road to the north and through the village of Liptovská Teplá we reach the village of **Lúčky (stop B7.9)**. From Lúčky, we drive to the state boundary at Trstená-Chyžne and back to Kraków along E77 (S7).

Introduction to the trip

Main topics

The field trip is an answer to rapidly growing interest in freshwater carbonates which, on the one hand, reflects their potential as a palaeoenvironmental archive and, on the other hand, is associated with the discovery of oil and gas in such kind of rocks in the South Atlantic. To some extent the trip follows the idea of similar ones organized during the former IAS conferences, that is Kraków in 1986 (Gąsiorowski *et al.*, 1986), Fukuoka in 2006 (Kano *et al.*, 2006), Alghero in 2009 (Capezzuoli *et al.*, 2009) and Zaragoza in 2011 (Vázquez-Urbez *et al.*, 2011), as well as some meetings especially devoted to such topics (Tata – 2004, Pammukale – 2005, Hull – 2008, and Abbadia San Salvatore – 2011). The field trip focuses on various types of freshwater carbonates in the Central and Internal Western Carpathians since this area abounds in travertines, tufas and caves highly decorated with speleothems. Travertines and tufas were, and still are, formed in different environmental and hydrological conditions. Their depositional history reflects geomorphic evolution of the area, tectonic events, climate changes, which in turn influenced palaeohydrological conditions, as well as even activity of prehistoric people. The above phenomena have been recorded also in speleothems since their

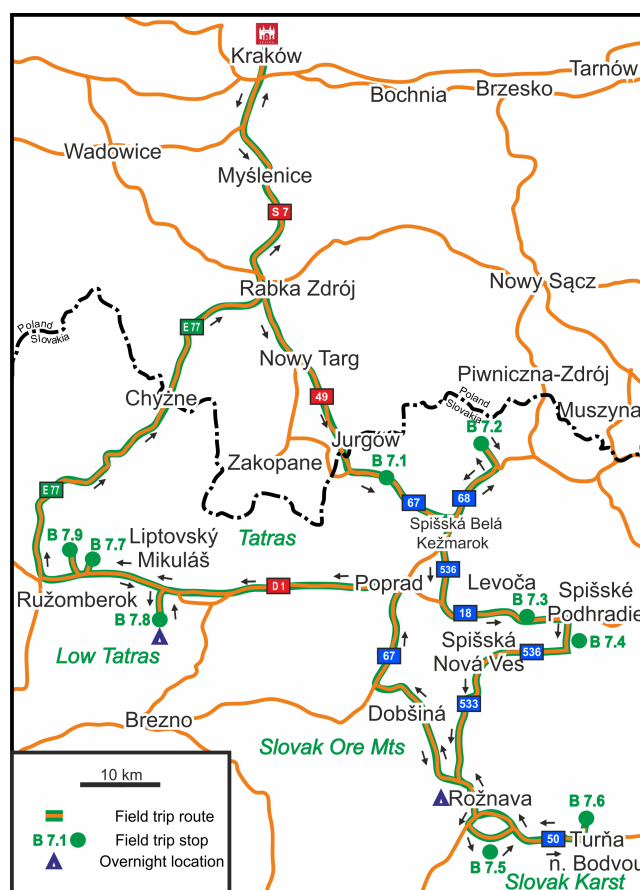


Fig. 1. Route map of field trip B7.

growth are under influence of similar environmental processes.

An overview of geology

Northern Slovakia represents an area of mountain massifs and intervening intramontane basins (Fig. 2). It occupies a part of the Carpathian chain; this part, from the geological point of view, belongs mostly to the Central Carpathians. The mountain massifs are predominantly built up of sedimentary rocks. They comprise carbonates, mostly Mesozoic, and especially Middle Triassic in age (Maheľ and Buday, 1968). Other Mesozoic rocks occur subordinately. The sedimentary rocks cover older, crystalline basements which constitute the cores of many mountain massifs. All the rocks in question were deformed in the Late Cretaceous time; several nappes were formed and pushed to the north as an effect of a crustal shortening resulting from the convergence between Adria-Africa and Europe (Plašienka, 2008). The Central Carpathians were subjected to erosion and denudation during Palaeogene times, to Early Eocene. Subsequently, tectonic subsidence created Central Carpathian Palaeogene Basins filled with deposits of a marine trans-

gression. A sequence comprising conglomerates, limestones, and a thick package (up to 3 km) of flysch-type rocks was laid down. This sequence is called the Central Carpathian Palaeogene. In Miocene, some parts of the region started to be uplifted; they presently form mountain massifs composed of Mesozoic rock and their crystalline basement. The intramontane basins are filled with the Central Carpathian Palaeogene, predominantly with carbonate-free flysch sandstones and mudstones. They cover the same Mesozoic rocks which constitute neighbouring uplifted mountain massifs. During the Middle–Late Miocene time, the Central Carpathians started to be under an extension regime due to a switch from an advancing to a retreating convergent system (Plašienka, 2008). At present, the area experiences an NW–SE compression and an NE–SW extension. Volcanic activity persisted in the Internal and Central Carpathians since Miocene (Eggenburgian) through Pleistocene (Lexa and Konečný, 1998).

The development of modern relief of northern Slovakia started in Miocene and it was associated with tectonic activity of the region. The surface fluvial drainage and underground karst drainage systems were created. The latter were drained by karst resurgences. Simultaneously, springs located along faults started to expel deeply circu-

lating water which migrated up across impermeable and insoluble rocks of the Central Carpathian Palaeogene. The recharge areas were located on the elevated massifs, whereas the springs were located mostly in the intramontane basins.

The geological history of the Inner Carpathians was slightly different. This region, including the Slovak Karst, experienced long lasting denudation, which started at the end of the Cretaceous (Mahel' and Buday, 1968). The area was drained towards the south, to the Pannonian Sea. In Late Miocene time (at the turn of Pannonian and Pontian), Slovak Karst was uplifted and tilted southward, which caused vigorous entrenchment of deep valleys and development of vertical caves (Gaál, 2008).

Travertines

Actively growing freshwater carbonates occur widely in northern Slovakia (Kovanda, 1971; Demovič *et al.*, 1972; Gradziński, 2010). They are situated near their pre-Holocene predecessors (Gradziński *et al.*, 2008b). The actively growing travertines are located near springs fed with deeply circulating waters, highly charged with CO₂ of crustal or even mantle origin (Hynie, 1963; Cornides and Kecskés, 1982; Lešniak, 1998; Povinec *et al.*, 2010). They display a very wide range of facies types; from those typical of travertines *sensu stricto* to those common in calcareous tufa. However, they have been genetically connected to deeply circulating water, or its mixture with water of shallow circulation. Therefore, for the sake of simplicity, in this guidebook they are named consistently as travertines, regardless of their texture and facies type (for terminological discussion see Pentecost, 2005; Pedley, 2009; Jones and Renaut, 2010; Capezzuoli *et al.*, 2014). Such an approach allows distinguishing between these deposits and calcareous tufas, which are common in southern Slovakia, but have originated in completely different geological and environmental setting.

Recent travertines compose spectacular morphological forms, as craters (some completely dried, some still filled with water; Ružbachy, stop B7.2) and cascades formed on inclined slopes and covered with microgours and microdams (Sivá brada, stop B7.3 and Bešeňová, stop B7.7). The growth rate of modern travertines reaches a few centimetres per year and definitely exceeds the growth rate of calcareous tufa in the same region (Gradziński, 2010).

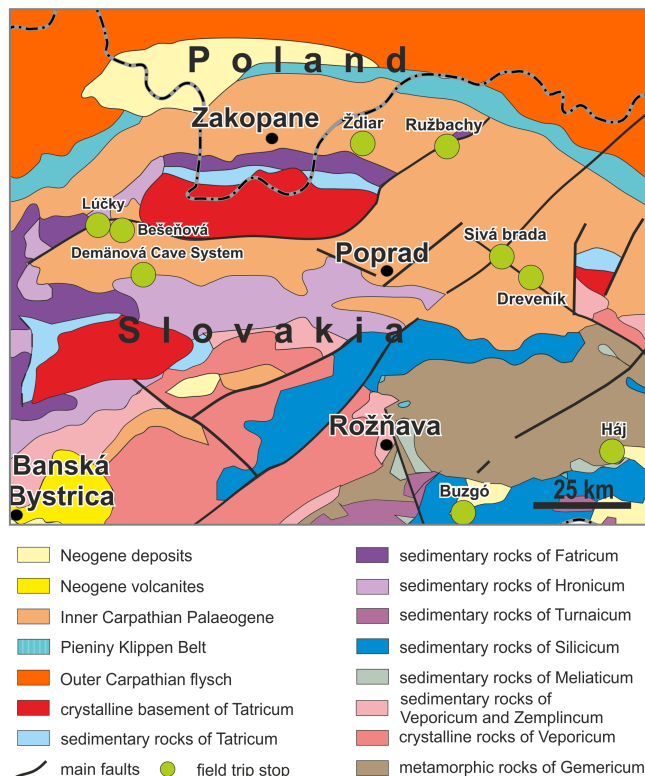


Fig. 2. General geology of the region (after Vozár and Káčer, 1996, modified) with location of the field-trip stops.

The growth of travertines is governed by chemistry of parental solution and the rate of CO₂ degassing, which decide upon effective precipitation of calcium carbonates (Pentecost, 2005; Pedley, 2009; Capezzuoli *et al.*, 2014). The physico-chemically stimulated quick crystallization of calcium carbonate affects the organisms in the milieu of travertine growth. Moreover, highly mineralized waters impede growth of many microorganisms and plants. Thus, the (micro)organisms seem to play subordinate, if any, role in the formation of many travertines. However, in some cases microbial and algal contribution is important, or even crucial, especially where deeply circulating and highly mineralized water mixes with water of shallow circulation (Lúčky, stop B7.9; Gradziński, 2010). Such settings are densely populated by cyanobacteria, algae, liverworts, mosses, and also higher plants. Some of them, especially those belonging to cyanobacteria and algae, can physiologically stimulate crystallization of calcium carbonate (see Rogerson *et al.*, 2008). Moreover, the organisms exert a substantial control on the texture of growing travertine. Comparison of isotopic composition ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) of modern Slovak travertines and isotopic composition of their parental water ($\delta^{18}\text{O}$) and dissolved total inorganic carbon ($\delta^{13}\text{C}_{\text{DIC}}$) reveals that travertines grow out of isotopic equilibrium, because of kinetic effects (see Kele *et al.*, 2008, 2011). This limits their usefulness for palaeoclimatic reconstruction.

Fossil, inactive travertines in northern Slovakia form buildups of various shape and dimensions (mounds, ridges, terraces). They reach the height of several dozen metres, whereas their lateral extent exceeds a few kilometres. They were formed, similarly to their modern counterparts, near springs fed with highly mineralized water of deep circulation. Their shape is controlled by the relief of basement surface, location of a potentiometric surface, CO₂ pressure, amount of feeding water and its distribution in travertine growth area, chemistry and temperature of this water, as well as the growth rate of travertine, which strongly depends on the above mentioned factors.

Travertine buildups display a variety of lithotypes, the same as those distinguished in Tuscany travertines by Guo and Riding (1998). The most common are crystalline crusts which developed on inclined slopes of the travertine buildups. At present, this lithotype is formed in cascades at Sivá brada (stop B7.3) and Bešeňová (stop B7.7) which are fed by a thin film of flowing water. Calcite

raft lithotype is associated with water ponded in small pools. Various types of rafts are being formed at Sivá brada. Fossil examples of crystalline crusts, calcite rafts as well as lithoclast travertine and coated bubble travertine compose an extensive, inactive travertine ridge, called Drevení (stop B7.4; Gradziński *et al.*, 2014). This ridge was affected also by postdepositional deformation, which resulted in fracturing and brecciation.

Although phytoclastic lithotypes and stromatolites are atypical of travertines, they commonly occur in mixing-water settings. The Lúčky site (stop B7.9) provides both, modern and ancient examples (Gradziński *et al.*, 2008b; Gradziński, 2010).

Pre-Holocene travertines in Slovakia are commonly regarded as originated during warm climate phases of Neogene and Pleistocene (interglacials). This notion is based on palaeobotanical, malacological and geomorphological data (Němejc, 1928, 1931, 1944; Petrbok, 1937; Ložek, 1957, 1961, 1964; Ložek and Prošek, 1957; Vaškovský and Ložek, 1972). However, it is intriguing that, in spite of favourable climatic conditions, the recent growth of travertines seems to be limited in comparison with the wide-spread occurrence and great lateral extent of the pre-Holocene travertine buildups.

Calcareous tufa

Conversely to northern Slovakia, fossil tufa outcrops and sites of modern tufa deposition are common in the Slovak Karst area, in southern part of the country. Kovanda (1971) and Kilík (2008) listed several tufa localities there. The Slovak Karst is a typical karst area with several plateaus built of Mesozoic carbonates. The plateaus are dissected by valleys with bottoms located approximately at elevation of 200–300 m, whereas the plateaus reach 400–800 m. Numerous crags and cliffs built of carbonate rocks occur on the plateau slopes. The plateau tops and north-facing slopes are forested with deciduous trees and their south-facing slopes are covered mainly by xerothermic grasslands and bushes. The plateaus are drained by extensive karst systems leading water to karst springs with average discharge up to 120 L/s (Jakál and Bella, 2008). The water is chiefly of the Ca–HCO₃ type and its mineralization exceeds 500 mg/L. The springs are located at the foot of the plateaus.

Active tufas precipitate near almost each spring in the Slovak Karst area (stops B7.5 – Buzgó and B7.6 – Háj). They

form tufa barrages and pools, cascades with tufa curtains, and oncoids (Kilík, 2008). The tufa depositional milieu is densely vegetated by cyanobacteria, algae, liverworts and mosses. These organisms are supposed to play an important role in tufa growth in this area (Gradziński, 2010).

Inactive tufas of Holocene age occur adjacent to active springs. The tufas in question are exposed by erosion and downcutting, in some places reaching to the Mesozoic bedrock. In a narrow valley setting tufas form a longitudinal fluvial depositional system with abundant barrages and inter-barrage areas. The former were built of moss, stromatolitic and phytoclastic tufa. The latter are filled with oncoidal and detrital tufas, including intarclastic ones. This tufa depositional system is governed by a limited ability of lateral stream migration. Conversely, tufas representing perched springline depositional system were formed below springs situated on the plateau slopes. Such a setting enabled omitting obstacles, impeded the formation of barrages and, hence, also dammed areas. Thus, cascades built of moss, stromatolites and phytoclastic tufa originated (Gradziński *et al.*, 2013).

In the Slovak Karst, the tufas grew especially vigorously in the Mid-Holocene, namely in the Atlantic and Sub-Boreal times (Gradziński *et al.*, 2013). Subsequently, they experienced substantial erosion and the streams incised down. This could be stimulated by deforestation by prehistoric humans or by late Holocene climate changes. The factors causing decline of the tufa growth ceased to operate, which is proved by the widespread formation of modern tufa in the Slovak Karst area.

Speleothems

Slovakia abounds in caves; more than 7,000 caves are registered (P. Holúbek, personal information, 2015). Some of them are accessible for tourists. Many caves are richly decorated with various types of speleothems. They represent inactive and actively growing ones, which was confirmed by analyses of feeding water.

The speleothems are a robust carrier of palaeoenvironmental information about palaeoclimate, tectonic/seismic activity, and geomorphic evolution of a karst region. Speleothems from Slovak caves have been recently extensively studied for these purposes. The analysis of speleothem growth frequency proved that they crystallized mainly in the warm periods of Pleistocene (Hercman, 2000). However, some far-reaching suggestions on

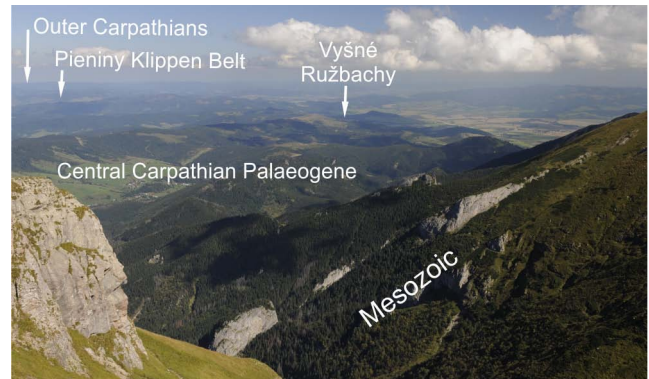


Fig. 3. Geological panoramic view of the area north of the Tatras. Mesozoic carbonates building the northern slopes of the Tatras (in the foreground) deep steeply to the north and plunge below the Central Carpathian Palaeogene. Pieniny Klippen Belt and the Outer Carpathians are visible in the background.

local climate anomalies were also put forward (Hercman *et al.*, 1997, 2008). Dating of speleothems shed some new light on the age of the extensive Demänová Cave System (in Slovak – Demänovský jaskynný systém; stop B7.8) and demonstrated that the lowest ‘levels’ of this cave are definitely older than it was postulated.

Stop descriptions

B7.1 Ždiar, viewpoint – outline of geology

(49°16'24" N, 20°13'44" E)

Leader: Michał Gradziński

This stop located on the edge of the Tatras enables us to familiarize with the general geological structure of the Central Carpathians (Fig. 3). The trip departed from Kraków, located on the boundary between the Kraków Upland (Carpathian foreland), a narrow segment of the Carpathian foredeep, and the Carpathians. Travelling to the south, we crossed several nappes of the Outer Carpathians, thrust to the north on the foreland basin filled with Miocene deposits. The nappes are composed of uppermost Jurassic to Early Miocene rocks, predominantly of flysch type. Next, we crossed the Pieniny Klippen Belt. It is a narrow zone stretching from Vienna (Austria) to Maramureş (Romania), regarded as a boundary between the Outer and Central Carpathians.

Between the Pieniny Klippen Belt and the Tatras, there is a hilly area clearly visible in the foreground (to the north). Its western part is called Podhale whereas the eastern one – Spisz (in Polish) or Spiš (in Slovak). It is

built of Central Carpathian Palaeogene rocks (mainly of siliciclastic flysch) attaining 3 km in thickness and forming an asymmetric syncline.

To the south, in the close vicinity, we can see steep, rocky slopes of the Tatras, in particular their easternmost part, called the Belianske Tatras with their highest summits Havran (2154 m) and Ždiarska Vidla (2148 m). The Tatras are the highest mountain massif of the Carpathians. They culminate in Gerlach (2655 m). They display morphological and geological asymmetry. Their northern slopes are built of several nappes, composed of Mesozoic rocks, predominantly carbonates, whereas their main ridge and, especially, its southern slopes are composed chiefly of various crystalline rocks. In Miocene the Tatras were uplifted along a prominent boundary fault on the south and south-east. Beside the Tatras, a small horst built of Mesozoic rocks is uplifted along this fault. The horst is engulfed by Palaeogene rocks, and due to this, it is called ‘an island’. The second stop of this trip is located there.

Palaeogene rocks, which originally covered the Tatras, have been completely eroded since Miocene times. The uplift and glacial erosion led to the formation of high-mountain, alpine-type relief of the Tatras. They underwent repeated glaciations during the Pleistocene. Several cave systems exist in the limestone parts of the Tatras. They discharge waters from the karst massifs in resurgences located at the foot of the mountains, but a substantial portion of water feeds confined aquifer (an artesian basin) located to the north of the Tatras, where Central Carpathian Palaeogene rocks act as confining beds. Such water naturally outflows in the area of the Ružbachy ‘island’ (Hanzel, 1987).

B7.2 Vyšné Ružbachy spa – Holocene and modern carbonates in the discharge zone of deep-circulation water

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Mineral springs in the village of Vyšné Ružbachy (in Polish – Wyżnie Druzbaki, in German – Oberrauschenbach, in Hungarian – Felsőzúgó) in the Spiš region have been mentioned in literature since 1549. The village has acted as a spa since the end of the 16th century (Potočná, 2007). It is worth mentioning that the first study on medical properties of water from Ružbachy was undertaken in 1635 and funded by Stanisław Lubomirski, the owner of Ružbachy and a prefect of the Spiš district which belonged to Poland at that time. The study was conducted by Jan Innocenty Petrycy, a professor of the Akademia Krakowska (now Jagiellonian University).

There exist nearly 20 natural and artificial (from drilled wells up to a few dozen metres deep) outflows of mineral water in the area of Vyšné Ružbachy. Distribution of the outflows is controlled by the geological structure of the area. Vyšné Ružbachy is located on the outskirts of the so-called ‘island’. It is an area where Mesozoic

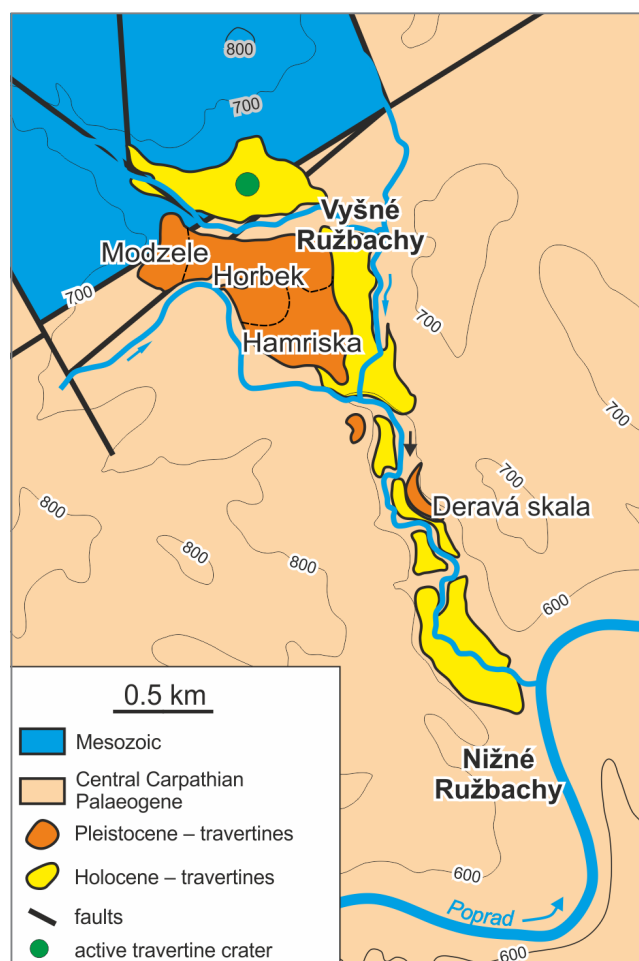


Fig. 4. Simplified geological map of the Ružbachy area (stop 2), after Ložek (1964) and Janočko *et al.* (2000).

rocks crop out from beneath the Central Carpathian Palaeogene siliciclastic rocks (Fig. 4). Mesozoic rocks belong to the Krížna Nappe) and represent the same rock series as in the Tatras, including the Belanské Tatras, which we observed from our first stop. Mesozoic rocks are bordered from the southeast by a fault with a throw of ca. 1.5 km. In the area of Vyšné Ružbachy it runs in the SW–NE direction, and it is the prolongation of the southern boundary fault of the Tatras, called the Sub-Tatric fault. Therefore, the Mesozoic ‘island’ of Ružbachy resembles to some extent the Tatras in Neogene time, when the Palaeogene cover only started to be eroded and Mesozoic rocks of the Tatras cropped out only locally. The Mesozoic rocks composing the north-eastern part of the Tatras and crystalline rocks of the Tatras core, which are hydraulically connected with them, act as the recharge area for outflows in Vyšné Ružbachy. The Middle Triassic carbonates, and Cretaceous and Eocene limestones constitute a confined aquifer, whereas Central Carpathian Palaeogene impermeable siliciclastics act as confining beds (Fendeková, 2002). The water ascends from beneath the confining beds partly along the Sub-Tatric fault and along smaller faults associated with it, and partly by the ‘island’ of Ružbachy which acts as a hydrogeological window.

Outflowing water has temperature between 17 °C and 24 °C. Total mineralization is within the range from 1.9–3.6 g/L. The water represents the HCO₃–SO₄–Ca–Mg type. Ca content ranges from 320 mg/L to more than 600 mg/L, HCO₃ from 1135 mg/L to more than 2000 mg/L. Carbonate equilibrium is controlled by the high CO₂ content, which varies between 360 and 1700 mg/L (Fendeková, 2002). H₂S is also present in small amounts (0.1–4.65 mg/L). High carbonate alkalinity of the water implies that carbonate rocks are being actively dissolved in the aquifer. This process takes place in elevated temperature. Therefore, one can presume that hypogenic caves are being formed by ascending water in deep substratum below impermeable Central Carpathian Palaeogene rocks. Such inactive caves have been recognized in northern slopes of the Tatras. One of them is Belianska Cave located in the E margin of the Tatras in the vicinity of the Sub-Tatric fault.

Spontaneous outgassing of CO₂ near the outflow zones enables quick and efficient calcium carbonate precipitation and formation of travertines. Nemejc (1931)

found several travertine localities near Vyšné Ružbachy. Subsequently, Ložek (1964) distinguished three generations of travertines (Fig. 4). The first one crops out in the hill called Modzele located west of the spa; it is visible in an abandoned quarry. It represents crystalline crust and coated bubble lithotypes; they are probably early Pleistocene in age (Gradziński *et al.*, 2008b; Rajnoga, 2009). Three localities – Horbek, Hamriska i Deravá skala – located to the S and SW of the spa represent the second generation. The abandoned quarry at Horbek is the largest locality, now hosting an open air gallery of modern sculpture. At Horbek, the travertines are composed of stromatolites, oncoids, phytoclastic and intraclastic lithotypes (Rajnoga, 2009). They most probably originated ca. 200 ka (Gradziński *et al.*, 2008b). The youngest travertine generation is Holocene in age, including recently growing travertines. This generation occurs within the spa and downstream over a distance of 2.5 km to the village of Nižné Ružbachy.

B7.2.1 Inactive travertine craters

(49°18'24" N, 20°33'27" E)

Inactive, sub-recent travertine craters are present ca. 250 m NW of the spa centre, at the edge of the forest. They are completely dewatered; grass, bushes and trees grow on their bottoms. The largest is 52 m x 32 m across, and 4.5 m deep. Exhalation of CO₂ is periodically active in one of them, hence it is called 'the death hole' (in Slovak – Jama smrti; Potočná, 2007). Dead animals from small creatures to birds and even foxes were found there. Weathered Upper Triassic rocks are visible to the north of the craters in a forest road.

B 7.2.2 Active travertine crater

(49°18'20" N, 20°33'37" E)

A crater spring (*kráter*) is the most widely known travertine site in Vyšné Ružbachy. The rim of the water-filled crater is circular in shape, 19 m across. The crater is up to 3.5 m deep. Water temperature is about 23 °C (Hynie, 1963). Mineralization of the water reaches 2364 mg/L and is dominated by HCO₃ (1384 mg/L), SO₄ (372 mg/L), Ca (393 mg/L), Mg (109 mg/L) and Na (42 mg/L) according to Fendeková (2002). The rim was formed by aggradation of travertine around an artesian spring. Its vertical

growth caused ponding of water, which in turn allowed further aggradation. Theoretically, this feedback mechanism can operate till the rim would reach a piezometric surface. The rim does not continue growing; water does not flow over it but is drained by an artificial culvert located in its southwest side and forms a stream.

B 7.2.3 Active travertine cascade

(49°18'16" N, 20°33'35" E)

A stream issuing from the crater runs SW and next SE and finally joins the Zálažný Stream which drains the Spišská Magura massif built of Central Carpathian Palaeogene rocks. Just near the crater, a part of water is captured for the spa purposes. Travertine is being precipitated almost along the whole course of the stream. This process is especially vigorous on two cascades, 5 m and 2.5 m high. Both cascades started to develop in 1998 when the stream course was artificially changed to its present position (ing. Maximilian Zavartkay – personal information, 2008). The amount of travertine precipitated on the cascades illustrates the rate and efficiency of its growth. Twice a year fresh travertine is removed to clean the streambed between the cascades in order to concentrate flow and prevent a lawn from flooding (ing. Maximilian Zavartkay – personal information, 2008).

Measurements on the upper cascade between November 2008 and June 2010 have shown that the growth rate of travertine reaches 1.73 mm per day in some places. This enormous growth rate of travertine was fully confirmed by a hydrochemical study. The water feeding the cascade was strongly mineralized (TDS between 1713 and 2286 mg/L) with the Ca content ranging from 239 to 407 mg/L. The content of Ca in water dropped abruptly from the top to the base of the cascade. The maximal drop equalled 99 mg/L. Bearing in mind the discharge of the feeding stream and chemistry of water, one may calculate the amount of calcium carbonate precipitated on the upper cascade per day; it ranges from 5.12 kg to the enormous value of 61.7 kg.

The isotopic composition of travertine at the cascade was studied between April 2008 and September 2009. The values of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of travertine vary from 4.6‰ to 7.5‰ and from -10.9‰ to -9.8‰ vs V-PDB, respectively. Isotopic composition of water at the cascade ranges from 0‰ to 10‰ (DIC), from -10.4‰ to -9.8‰

($\delta^{18}\text{O}$) and from -72‰ to -77‰ (δD). No correlation has been detected between isotopic composition of travertines and air temperature during the year. Uranium and polonium activities vary from 0.001 Bq/g and 0.006 Bq/g and 0.003 Bq/g and 0.020 Bq/g, respectively.

B7.3 Sivá brada – Active travertine cascade, modern calcite rafts

(49°0'22" N, 20°43'22" E)

Leaders: Michał Gradziński, Marek Duliński¹, Jacek Motyka, Janusz Baryła², Joanna Czerwik-Marcinkowska³, Mariusz Czop⁴, Teresa Mrozińska³

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The hill known under a meaningful name Sivá brada (Grey Beard) is located in the Hornád Intramontane Basin, between two historic towns, Levoča to the west and Spišská Kapitula to the east (Fig. 5). The whole area belongs to the historic region called Spiš (in Polish – Spisz, in German – Zips, in Hungarian – Szepes). This area, or its parts, belonged to Hungary, Poland, Austrian Empire, Austro-Hungarian Monarchy and Slovakia

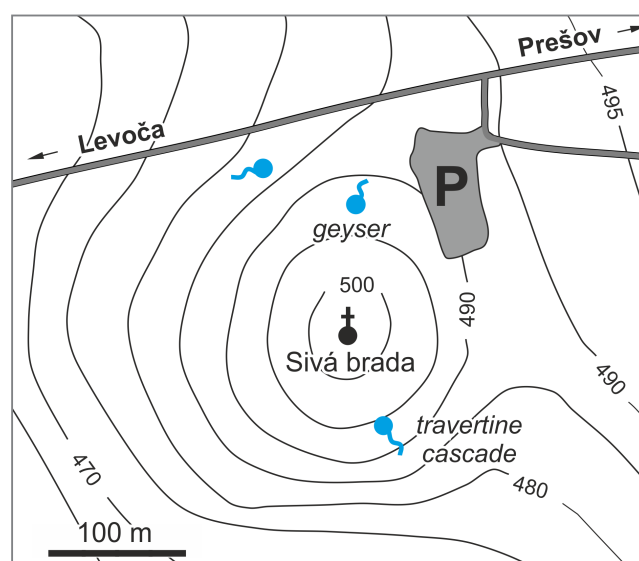


Fig. 5. Location of Sivá brada travertine mound (stop 3).

during its rich history. The hill is well visible from the road 18 Poprad – Prešov, and from the newly constructed motorway. A small chapel stands on the hill's summit (506 m). Springs with travertine precipitation are present on the hill slopes. A CO₂ exhalation is present in a small hollow by the summit chapel. The hill is protected as a nature monument.

Sivá brada is built of travertines and is a travertine mound, similarly as the nearby hills: Pažica, Hradný vrch, Ostrá hora, Dreveník, Sobotisko. The age of Sivá brada is unknown; Maglay and Halouzka (1999) postulate a Pleistocene, whereas Kovanda (1971) suggests a Holocene age. Clastic rocks of the Central Carpathian Palaeogene, up to ca. 1000 m thick, overlie the travertines. The travertines are underlain by Mesozoic carbonate rocks (Gross *et al.*, 1999). Sivá brada is located at the intersection of faults which provide suitable migration paths for ascending water from the carbonate bedrock to the surface. Recharge area for waters outflowing in Sivá brada and its vicinity is most probably located to the south, in the elevated area of the Slovak Ore Mountains (in Slovak – Slovenské rudohorie) – a mountain chain built of the same rocks as those underlying the Hornád Intramontane Basin.

The summit of Sivá brada provides an excellent view over the Hornád Intramontane Basin and its vicinity. To the east, several travertine mounds are visible. The closest one, with a small chapel perched on the top, is called Pažica. In the background, the travertine ridge stretches north–south. It comprises three hills, Hradný vrch (hidden behind Pažica), Ostrá hora, Dreveník (with the our next stop). An extensive massif of Branisko is located on the horizon. It is composed of Palaeozoic crystalline rocks and their Mesozoic sedimentary cover, thus it has similar geology to the Tatras. To the south, a hilly area is visible in the foreground. This is the Hornád Intramontane Basin carved in clastic rocks of the Central Carpathian Palaeogene. Farther to the south, lies the forested area of the Slovak Ore Mountains.

A few dozen metres south from the top of Sivá brada and 8 m below it, there is a small travertine cascade. It is fed by ascending water in the two pools located one over another. The lower pool is 5 m x 3.5 m in size and its depth reaches 20 cm. It was artificially created as a small puddle for spa purposes (Hynie, 1963). The

discharge of the springs changes with time, one or the other being more active. Water flows down the lower pool, spills over and forms a thin film feeding a travertine cascade. The cascade with microdams and micropools is inclined at an angle of 10°. A similar cascade lies between the pools.

The water is of the Ca–Mg–Na–HCO₃–SO₄ type with TDS reaching 6.9 g/L. Tests have shown absence of tritium, which means that the water as a whole recharged before 1952, that is before the beginning of thermonuclear tests in the atmosphere. It is charged with CO₂ of deep origin, whose content reaches 2343 mg/L. Partial pressure of CO₂ being in contact with water feeding the pool, calculated for the saturation point with respect to CaCO₃, equals approximately 6 atm. Mean Ca content is 829 mg/L whereas HCO₃ content is 3804 mg/L. Considering the pH and temperature of water, the calculated saturation index with respect to calcite is ca. 0.71–1.03 and ca. 0.87–1.42 in the lower pool and at the lower cascade, respectively. Such chemical composition of water results in fast and efficient precipitation of calcium carbonate in the pools, especially in the lower one, and on the surface of the cascades.

Two different sedimentary sub-environments of travertine growth are present here. The first are the pools with nearly stagnant water, whereas the cascades with microdams and micropools fed with a thin film of flowing water are the second one. The water feeding the cascades is distributed by 'self built canals', bordered by natural levees. Some clumps of vascular plants occur within a lower travertine cascade, especially in its inactive parts. The plant community comprises *Plantago maritima*, *Triglochin maritima* and *Centaureum littorale* which are classified as halophytes typical of a sea shore (e.g., Košťál, 2011).

Precipitation of calcium carbonate is reflected in the chemical evolution of water along its flow path below the lower pool. CO₂ and Ca contents decrease, whereas pH and saturation index with respect to calcite increases downward.

Calcite rafts are the most common type of carbonate precipitates which are formed in the lower pool. Two types of rafts have been recognized: paper-thin rafts and composite rafts. Both are composed entirely by calcite, although the water is supersaturated also with respect to aragonite. The first group is represented by rafts with

up to several micrometers thick. They consist of flat, ultra-thin film of calcite which does not display visible crystals. The lower side of such a film is overgrown by crystal aggregates. They exhibit hemispheric, barrel or dumbbell morphology and are composed of radially oriented needle-shaped subcrystals. This suggests that the crystals grow under disequilibrium conditions (cf. Jones and Renaut, 1995). The paper-thin rafts float freely due to surface tension of water and cover the major part of the pool surface.

The composite rafts float partly submerged in the pool water. They are overgrown by calcite on the upper and lower sides. Commonly, the rafts form piles composed of several individual paper-thin rafts, lying one on another. The rafts are partly broken and arranged subparallel. They do not sink completely because they are supported by an extensive organic buoyant mat, composed mainly of cyanobacteria, diatoms and their extracellular polymeric secretions. Sixteen algal taxa forming a mat have been distinguished. The most common are filamentous cyanobacteria of genus *Phormidium* and diatoms *Achnanthes minutissima* and *Navicula gregaria*. Diatoms are also attached to the rafts. Although some cyanobacterial sheaths, diatom mucilage and frustules are calcified, the majority of algae are not covered with calcium carbonate, which is proved by observation under scanning electron microscope. It shows that the algae do not participate in this process actively, but act only as a substrate for growing crystals. However, the crystals seem to grow more readily on mineral substrate, that is on sunken calcite rafts.

Outgassing of CO₂ is the principal process responsible for effective precipitation of calcite crystals and raft formation. Kinetics of outgassing depends also on seasonal changes of water temperature in the pool. This causes more efficient growth of the rafts in late spring, summer and the beginning of autumn. The influence of evaporation seems to be insignificant for the raft growth, although this process slightly modifies isotopic composition of the water in the pool.

Although the algae do not contribute actively to calcite precipitation, their role is important in the formation of the rafts since they buoyantly keep the rafts near the surface, where the effects of degassing are strong and supersaturation is high enough for calcite precipitation. Since the process occurs below the water

surface, the crystals grow on both bottom and top sides of rafts. Calcite rafts are extremely susceptible to destruction. It particularly concerns the paper-thin ones. Even loading by such a small object as a pollen grain causes deformation of their surfaces. Both, the paper-thin and the composite rafts are destructed by rain; however, the latter obviously have highest potential for preservation. They are a dominant component littering the bottom of the pool. Comparison of the present topography of the pools with an archival photography taken in the middle of the last century (Hynie, 1963, fig. 73 – 1) suggests that the lower pool was filled by deposits several dozen centimetres thick.

The cascades are built of white, laminated travertine composed of calcite crystals ranging from a few dozen to a few hundred micrometres. Observations suggest that vertical growth of the lower cascade exceeds 2.5 cm per year in the zones of active water flow.

Small spheres, up to 1 cm in diameter, composed of calcite crystals, form in micropools. They are encrustations on the surfaces of gas bubbles. In summer, microdams are roofed by a calcite film. It is similar to paper-thin rafts, originated over the water filling the micropools. The film has a slightly convex-up shape due to a meniscus head effect. It is very fragile and can be easily destructed. Debris of the calcite film accumulates in the micropools. Piles of fossil calcite film fragments most probably form puff pastry like fabrics (*sensu* Gandin and Cappezzuli, 2014).

The recently growing travertines (both rafts and cascades) show values $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ from +10.4‰ to +12.7‰ and from –10.0‰ to –6.8‰ vs V-PDB, respectively. Comparison of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analytical results with analogous values calculated for the isotopic equilibrium conditions indicate that the travertines grow in isotopic disequilibrium. The lack of equilibrium results from strong kinetic effects related to the fast outgassing of CO₂ to the bulk atmosphere.

An active ‘cold geyser’ on the northern slope of Sivábrada is clearly visible from the car park at the north-east foot of the mound. It is fed by a 135 m deep drill-hole (Jetel, 1999) and its spout water due to CO₂ pressure. It erupts irregularly. Although at present the eruption height is a few dozen centimetres, it reached more than 10 m (Kovanda, 1971). Water represents the Ca–Mg–Na–HCO₃–SO₄ type (Jetel, 1999). The geyser orifice is enveloped by

actively growing travertine. However, it is constantly destructed by trampling since it is a kind of tourist spot.

B7.4 Dreveník ridge, Žehra quarry – Facies and anatomy of an inactive travertine ridge

(48°58'55" N, 20°46'28" E)

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Dreveník is the southernmost part of a meridionally elongated ridge with the ruins of a medieval castle, Spišský hrad, perching on its northernmost part called Hradný vrch (Fig. 6). The castle was erected in 11th and 12th centuries; it covered approximately 4 hectares. The castle was built of local stone, predominantly travertine. Dreveník rises up to altitude of 609 m and is 2.5 km long and 1 km wide (Tulis and Novotný, 2008). It is one of the biggest travertine buildups in Central Europe. Travertine is up to 80 m thick and it overlies flysch deposits of the Central Carpathian Palaeogene, which plays a hydrological role of the confining bed (Fig. 7; Gross *et al.*, 1999). The orientation of the Dreveník ridge follows a fault in the basement. The age of travertine is considered to be Pliocene–Pleistocene based on the palaeobotanical (Němejč, 1944) and palaeontological finds (Holec, 1992; Tóth and Krepmaská, 2008) but it has not been defined precisely yet. Travertine is intensively eroded. It has undergone gravitational processes and karstification (Fusggänger, 1985; Tulis and Novotný, 2008). Many post-depositional features are present there, including widened fissures, cracks, karren, caves filled with secondary deposits of different origin (Wróblewski *et al.*, 2010).

Fifteen quarries exist at various points of the Dreveník ridge (Tulis and Novotný, 2008). The largest one, continuously active and used by the Euro Kameň company, is located on the western slope of the ridge. An abandoned quarry on the southern side of the Dreveník ridge, named Žehra quarry, will be visited

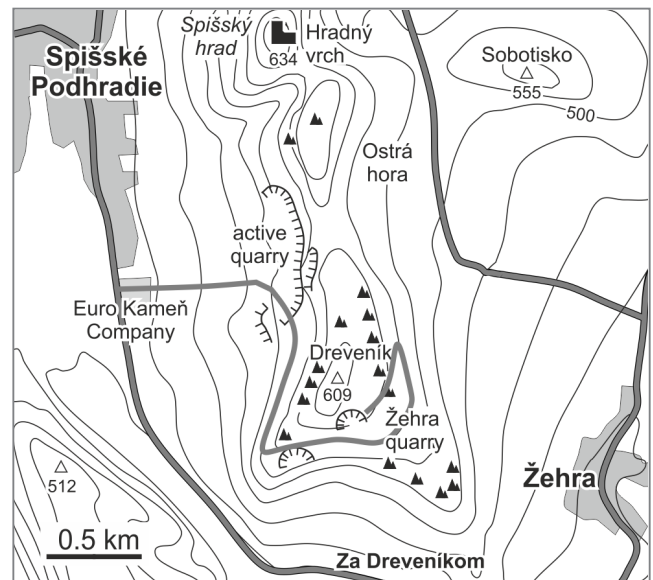


Fig. 6. Location of Žehra quarry (stop 4).

during the trip. Visiting of the quarry is allowed; however, the road used during the trip is closed and driving needs permission of an owner of the Euro Kameň company. The quarry may be accessed on feet from public roads around the hill and from Spišský hrad. The quarry was exploited from the seventies to the nineties of the last century (Tulis and Novotný, 2008). The quarry has eight levels. Travertine was mined by using saw ropes, hence the quarry walls consist of flat surfaces up to 1.8 m high several meters long. This allows us to observe various travertine lithotypes (*sensu* Guo and Riding, 1998), their spatial relationships and post- and syn-depositional features (Gradziński *et al.*, 2014).

Layered light grey and yellowish crystalline crusts travertine is the dominant lithotype in the quarry (Fig. 8A, B). Individual layers are between 0.5 cm and several centimetres thick. This facies has a very low porosity. The layers are inclined usually between 20° and 30°, rarely up to 70°. Crystalline crust travertine is composed of fan-shaped sparry calcites which grew from a thin water film on the ridge slope.

Lithoclast travertine is the second most common lithotype in the quarry (Fig. 8A). It forms four lobe-shaped bodies of different size, interbedded with crystalline crusts and other lithotypes and 0.2 m to 6 m thick. Lithoclast travertine is composed of subangular and angular clasts of different size (from a few millimetres to several decimetres) composed mainly of crystalline crust travertine, subordinately of vein calcites and micritic

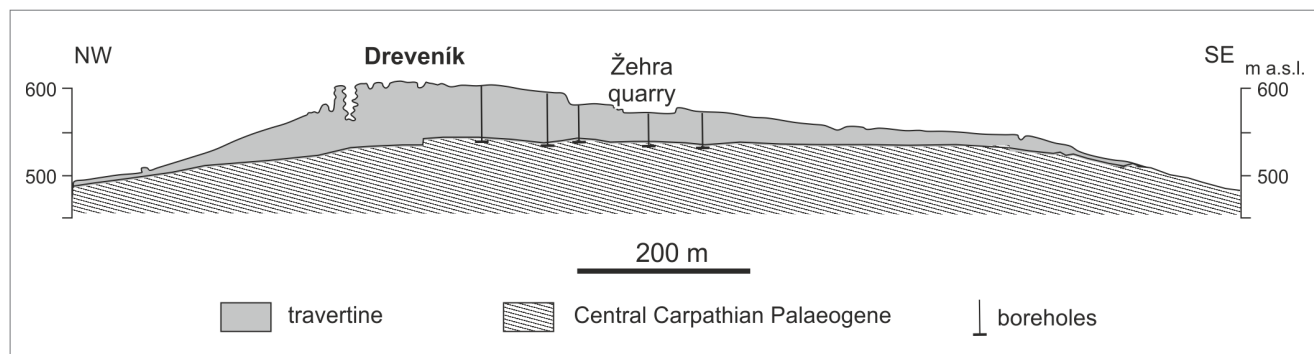


Fig. 7. Cross-section through Dreveník travertine ridge, after Tulis and Novotný (2008), modified.

travertine. The clasts were transported and deposited due to episodic rock falls (Gradziński *et al.*, 2014).

Fine-grained lithoclast travertine is another lithotype recognized in the Žehra quarry. It forms two lenticular bodies, up to 0.8 m thick. The lithoclasts are built mainly of crystalline crust travertine. Individual clasts reach up to several centimetres in size. Fine-grained lithoclast travertine is more porous than lithoclast travertine. It originates from erosion on the external parts of the travertine ridge and redeposition of its small fragments downslope in periods of limited spring water supply. It corresponds to hillwash breccia (*sensu* Pedley, 2009).

Coated bubble travertine and paper-thin raft travertine occur subordinately in the Žehra quarry (Fig. 8B, C). They usually form horizontal or sub-horizontal layers within crystalline crust travertine. The individual bubbles are vertically elongated and are up to 3 cm high. They grew on small ponds formed on the ridge slope. Paper-thin raft travertine comprises cemented thin rafts (up to 2 mm of thickness) accumulated in small pools or they fill empty spaces within intraclast breccia (Fig. 8B).

Several features including intraclast breccia, ground and long fissures, karst cavities related to syn- and post-depositional processes have been described in the Žehra quarry (Wróblewski *et al.*, 2010; Gradziński *et al.*, 2014).

Intraclast breccia is a product of syn-depositional brittle deformations. It occurs below the bodies of lithoclast travertine. Intraclast breccia consists of crystalline crust travertine clasts (Fig. 8B). The clasts are up to several decimetres in size. Individual clasts display limited displacement and they fit one another like parts of a jigsaw puzzle.

Long fissures constitute sub-vertical and vertical cracks in travertine (Fig. 8C). Their vertical extent exceeds 20 m. They are up to 0.7 m wide and are mainly filled

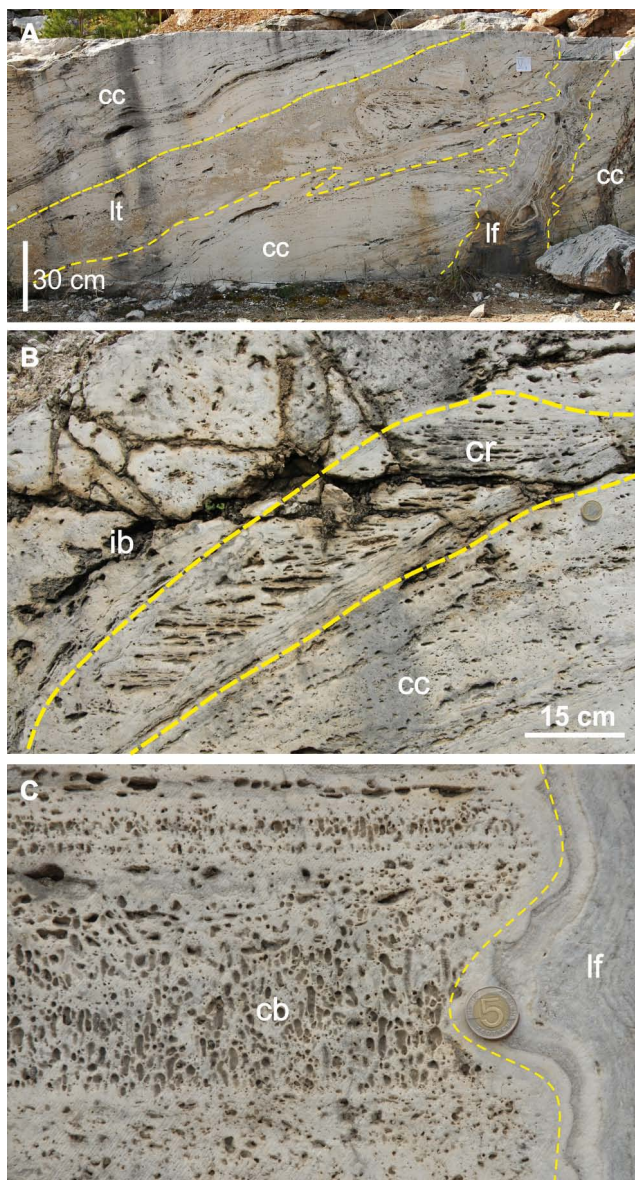


Fig. 8. Selected travertine lithotypes in Žehra quarry (stop 4). A) Lithoclast travertine (lt) interfingers between crystalline crust travertine (cc) and, younger long fissure (lf) is filled with vein calcite; B) Calcite rafts (cr) filling ground fissure between crystalline crusts (cc) and intraclast breccia (ib); C) Coated bubble travertine (cb) cut by long fissure (lf) with vein calcite.

with vein spelean calcites of phreatic origin (Gradziński *et al.*, 2014).

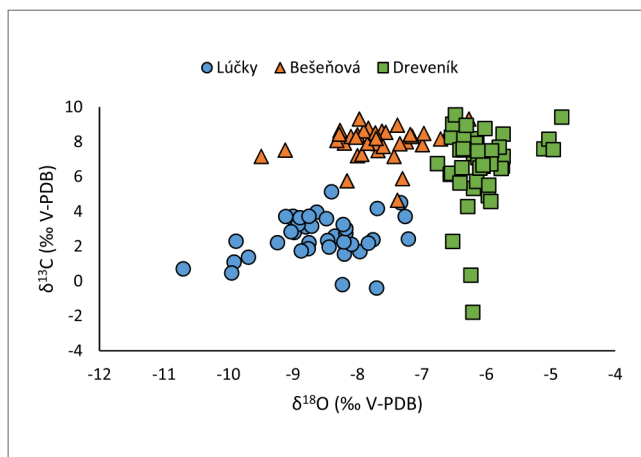


Fig. 9. Carbon and oxygen stable isotope composition of pre-Holocene travertine from Dreveník, Bešeňová and Lúčky, after Gradziński *et al.* (2008b, 2014) and unpublished data.

Ground fissures represent a group of cracks which are filled with several types of travertine, including crystalline crust travertine, lithoclast travertine and calcite rafts. They are shorter and less open than long fissures.

The travertine at Dreveník is of Pliocene age. This age is supported by findings of a pollen assemblage typical of the Central Carpathian Neogene. This assemblage includes pollens of *Carya* sp. and *Tsuga* sp. Palaeomagnetic study proved that it is normally magnetized. These data collectively suggest that the travertine was formed between 3.4 and 2.48 Ma.

The Dreveník travertine represents smooth slope facies (*sensu* Guo and Riding, 1998). Travertine was fed with highly-mineralized waters of deep circulation flowing out along the faults cutting the Central Carpathian Palaeogene rocks. Their $\delta^{13}\text{C}$ values are between -2.0‰ and $+9.6\text{‰}$ V-PDB, indicating that the waters were charged with CO_2 of geogenic origin (Fig. 9; Gradziński *et al.*, 2014). The Dreveník travertine can be regarded as a fossil analogue of the travertine ridges known from Turkey (e.g., Mesci *et al.*, 2007; Piper *et al.*, 2007; De Filippis *et al.*, 2012), Tuscany (Brogi and Capezzuoli, 2009; Pedley, 2009) and USA (De Filippis and Billi, 2012). The deposition of lithoclast travertine, as well as origin of its deformation is interpreted as a result of seismic shocks (Gradziński *et al.*, 2014). The seismic events exerted indirect control over the facies distribution. Local hydrological conditions changed after the shocks, which led to ponding of water and formation of paper-thin raft travertine.

B7.5 Buzgó stream – Deposition of modern fluvial calcareous tufa

(48°37'02" N, 20°35'04" E)

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Buzgó is one of the sites of present-day calcareous tufa sedimentation in the Slovak Karst. It is located in southern peripheries of the village Krásnohorská Dlhá Lúka, ca. 5 km SE of Rožňava (Fig. 10). The area is protected as a part of a national park. The site is named after the stream flowing over the fluvial terrace of the Čremošná river (Gaál, 2008). The Buzgó stream debouches in the resurgence of Krásnohorská Cave (in Slovak – Krásnohorská jaskyňa) which originated as the underground drainage of the northern part of the Silica Plateau (Orvan, 1980). The resurgence is located at the altitude of 316 m, at the foot of the plateau, approximately 1 km southeast of the village centre (Bella, 2008; Haviarová *et al.*, 2012). Its discharge ranges from 6 to 1300 L/s (Stankovič *et al.*, 2005).

Freshwater carbonates form in surface and subsurface (cave) section of the Buzgó stream. In the cave section of the stream, freshwater carbonates form subaqueous flowstones which are present on the streambed (Fig. 11A) along the distance of approximately 150 m upstream from the resurgence. Calcareous tufas are being formed in the surface section of the stream (Fig. 11B). Tufas commence to accumulate in the upper

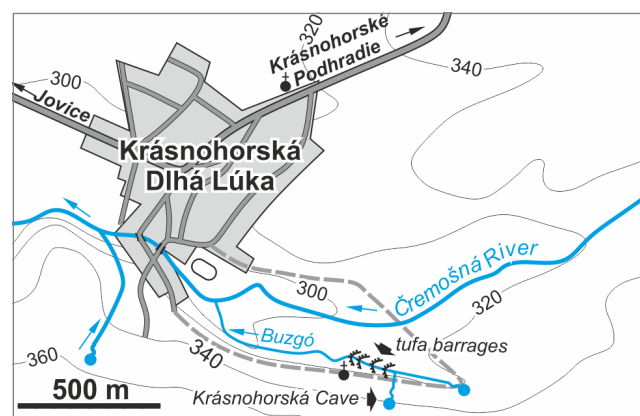


Fig. 10. Location of Buzgó resurgence (stop 5).



Fig. 11. Buzgó stream. **A)** subaqueous flowstones in subsurface section of the stream, the tablet arrowed was exposed in the stream during 128 days; **B)** multi-stepped tufa barrages in surface section of the stream.

reaches of the stream, approximately 20 m downstream of the resurgence. They construct multi-stepped barrages (Fig. 11B). Tufa covers an area up to 1 ha (Stanković *et al.*, 2005). The tufas represent perched springline facies association (*sensu* Pedley *et al.*, 2003), including such facies as stromatolitic tufa, phytoclastic tufa, moss tufa and oncoidal tufa. Stromatolitic tufas are the dominant ones. They are laminated and mainly composed of calcified cyanobacterial filaments. Stromatolitic tufas form barrages orientated perpendicular to the flow direction. Locally stromatolitic tufas interbed with layers and lenses of phytoclastic and moss tufas. Oncoids occur in small pools formed upstream barrages. Oncoids are sub-angular whereas their shape is cylindrical and spherical. They range from 1 to 15 cm across. Their formation is probably related to biological activity of some microorganisms in mid-energetic conditions.

Water of the Buzgó stream represents the Ca-HCO₃ type. Its TDS ranges from 367 to 836 mg/L (mean value = 565 mg/L). Other parameters are as follows: mean pH – 7.73, mean HCO₃ content – 367 mg/L, whereas mean

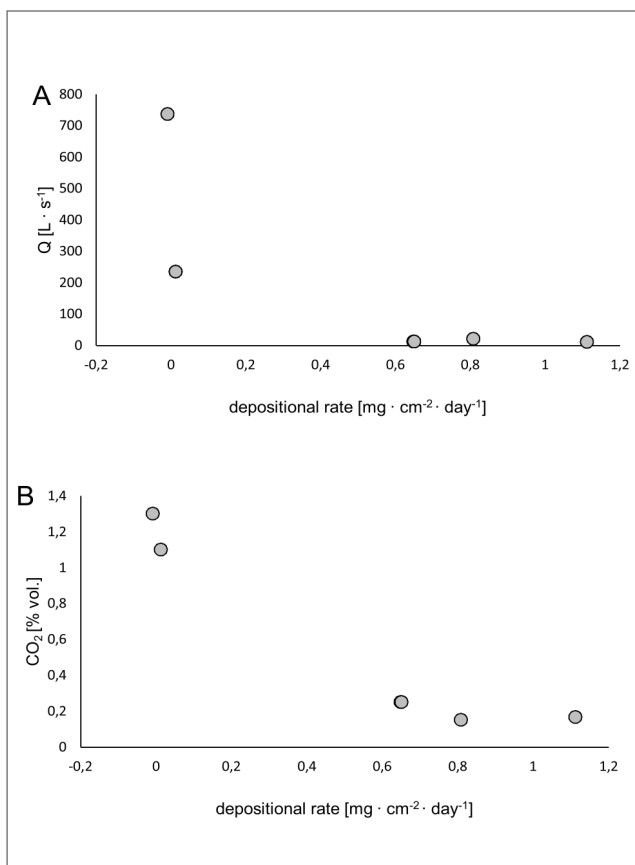


Fig. 12. Depositional rate of calcite measured on plates located 40 m downstream the resurgence depends on spring discharge (**A**) and mean CO₂ concentration in cave atmosphere (**B**).

Ca content – 124 mg/L. Waters are predominantly super-saturated with respect to calcite during the year (mean value SI = 0.68). The temperature is quite constant and oscillates around 9.5 °C.

Seasonal observations (water chemistry, precipitation rate on and CO₂ concentration in cave atmosphere) were conducted between August 2010 and September 2012. They show that precipitation of calcium carbonate is almost continuous all year round. The highest depositional rate was noted in the winter–spring season of 2012 (1.11 mg · cm⁻² · day⁻¹). Changes in the depositional rate of tufa positively correlate with fluctuation of groundwater levels (Fig. 12A) and CO₂ concentration in the cave atmosphere (Fig. 12B). This suggests that depositional processes are significantly modified by groundwater level fluctuations. Low groundwater levels in the subsurface segment of the Buzgó stream make ventilation of the karst system more effective, which in turn affects precipitation of calcium carbonate. This fully confirms the opinion formulated by Kano *et al.* (2003) and Kawai *et al.* (2006) on the influence of subsurface processes on tufa growth.

B7.6 Háj Valley – Facies and depositional history of Holocene fluvial calcareous tufa

(48°38'23"N, 20°50'57"E)

Leaders: Michał Gradziński, Joanna Czerwik-Marcinkowska, Helena Hercman, Martyna Jaśkiewicz¹, Jacek Motyka, Teresa Mrozińska, Stanisław Szczurek^{1,2}

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The Háj Valley (in Slovak – Hájska dolina) is narrow and it is incised to a depth ca. 150–200 m into the surrounding carbonate plateau built up of Triassic carbonates of the Silica Nappe. The head segment of the valley is carved into the Bôrka Nappe which is mainly composed of metamorphic rocks (Mello *et al.*, 1996). The trip will visit the middle segment of the valley. The coordinates given above point to a small car park on the west side of the road. The whole area is protected as a part of a national park.

Tufa extends over the distance of 900 m up the valley from the village of Háj (Fig. 13; Gradziński *et al.*, 2013). The stream flows down the valley from a series of karst springs. It forms four waterfalls, each a few metres high. Recently tufa is being formed in the stream as small barrages and scenic curtains hanging down from the heads of the waterfalls (Gradziński, 2010). The water is of the Ca–HCO₃ type; its TDS varies from 315 to 421 mg/L, whereas SI is between 0.17 and 1.1.

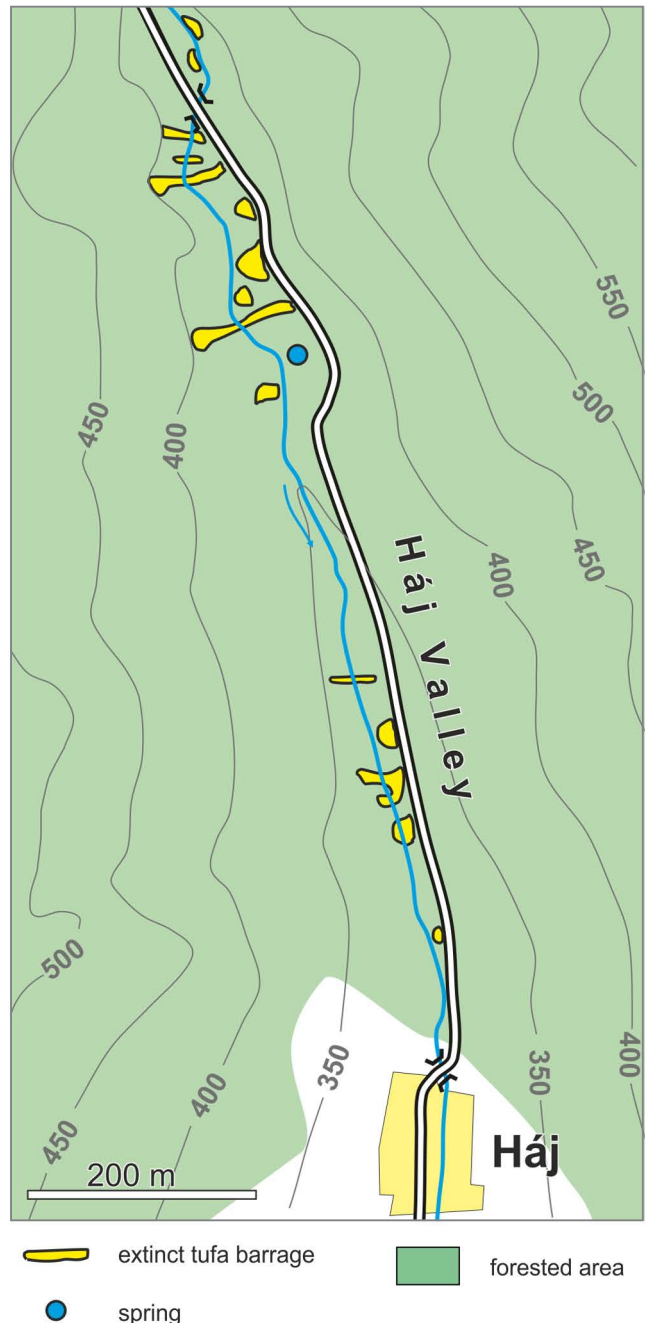


Fig. 13. Location of Holocene tufa barrages in the Háj Valley (stop 6; after Gradziński *et al.*, 2013, modified).

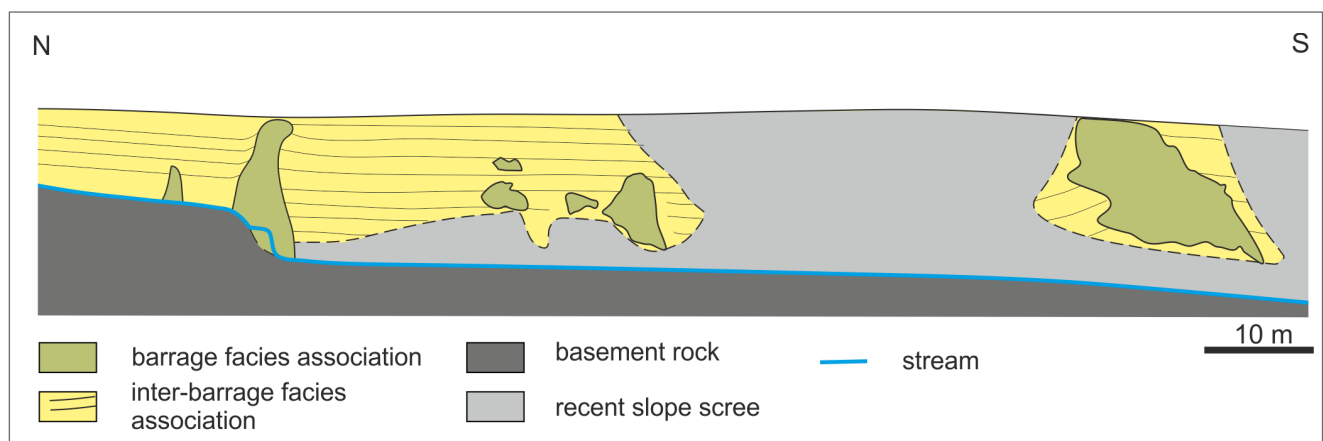


Fig. 14. Spatial distribution of barrage facies association and inter-barrage facies association in the upper segment of the Háj Valley, cross-section, after Gradziński *et al.* (2013), modified.

Inactive tufa partly fills the bottom part of the valley. It is incised, in many places even to its Mesozoic bedrock. Tufa sections crop out on terrace risers. The present relief of the valley clearly reflects the distribution of facies type associations of inactive tufa. The relatively hard and resistant tufa forms jumps and constrictions in the longitudinal profile of the valley, whereas wider valley segments are carved in loosely cemented tufa (Fig. 14; Gradziński *et al.*, 2013).

Tufa in the Háj Valley was noticed by Kormos (1912). Petrboč (1937) and Ložek (1958) carried out detailed malacological studies whereas Němejc (1936, 1944) studied plant fossils.

Tufas in the Háj Valley correspond to a longitudinal fluvial system (Gradziński *et al.*, 2013). It comprises barrage and inter-barrage facies association (Fig. 14). Barrages are composed of moss tufa which comprises three-dimensional, reticulate fabrics built of calcite-encrusted moss stems and leaves. Phytoclastic tufa, formed by calcite encrustation on plant fragments which

are preserved as empty moulds, co-occurs with moss tufa. Stromatolitic tufa is the third facies building barrages. Seventeen inactive barrages, up to 12 m high, occur in the Háj Valley (Fig. 13, Gradziński *et al.*, 2013). They are ponding water in the upstream reaches of the valley, where inter-barrage facies association was laid down. It constitutes oncoidal and intraclastic tufa which display grading and cross-bedding. Thus, this facies association originated in flowing water not in stagnant ponds. It corresponds to the ‘braided fluvial model’ of Pedley (1990) or ‘free flowing water channel-filling sequences’ of Vázquez-Urbez *et al.* (2012).

Tufas are locally covered with colluvial breccias composed of angular, poorly sorted clasts of Triassic carbonates (Gradziński *et al.*, 2013). They are bound by stromatolitic coatings, whereas intraclastic tufa acts as matrix. The clasts derived from upper slopes of the valley and are a rockfall or rockslide deposits introduced into the tufa-depositing system.

Radiocarbon dating proves that tufa grew in Mid Holocene time, that is during the Atlantic and Sub-Boreal intervals (ca. 7.5–3.5 ka BP; Fig. 15A; Gradziński *et al.*, 2013). The growth rate can be estimated at around 2.5 cm per year. Redeposition from the slopes and deposition of coarse-grained colluvium preceded tufa erosion and the stream incision (Fig. 16B). Such a phenomenon is clearly visible in many European tufa complexes. Goudie *et al.* (1993) coined a term ‘late Holocene tufa decline’ and discussed several factors that may have been responsible for it. In the Slovak Karst case, erosion is hypothesized to have been stimulated by deforestation caused by prehistoric humans (Gradziński *et al.*, 2013); however, other reasons cannot be unequivocally excluded.

The factors responsible for tufa erosion must have ceased to operate; modern tufa grows vigorously in the Háj Valley. The study conducted in 2002 and 2003 documents the tufa growth rate. Tufa, which constructs a curtain constantly flushed with water in the upper segment of the valley, is precipitated at a rate up to $0.3 \text{ mg} \cdot \text{cm}^{-2} \cdot \text{day}^{-1}$ (Gradziński, 2010). The curtain is constructed of encrusted cyanobacteria (*Phormidium favosum*, *Ph. tenue*, *Oscillatoria limosa*), algae (*Cladophora glomerata*, *Vaucheria* sp.) and mosses (*Brachythecium rutabulum* – R. Ochrya, pers. inf., 2003).

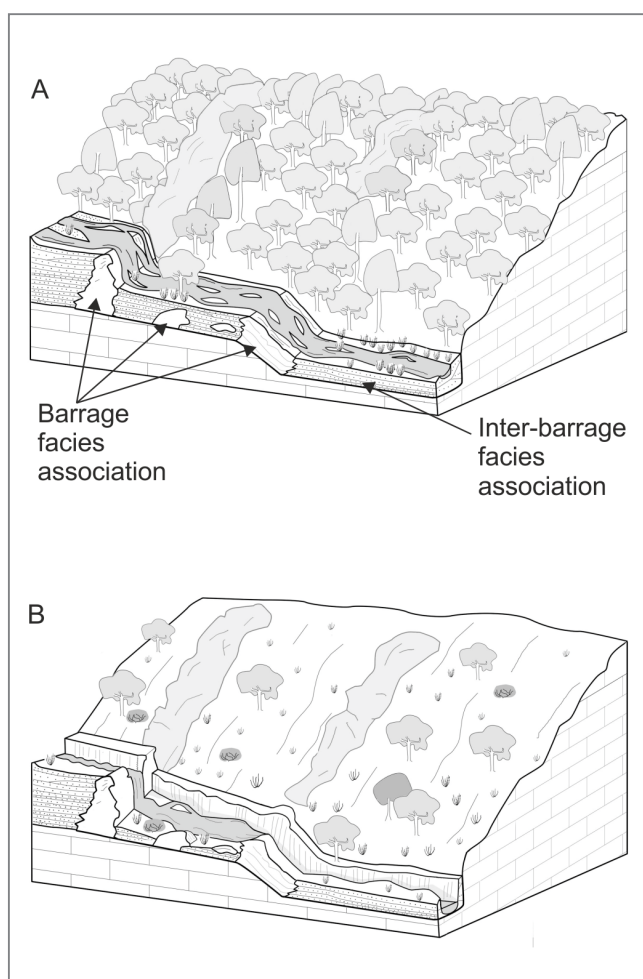


Fig. 15. Deposition (A) and subsequent erosion (B) of tufa in the Háj Valley (after Gradziński *et al.*, 2013).

Stop B7.7 Bešeňová – Pleistocene and recent travertines

Leaders: Michał Gradziński, Pavel Bella

The village of Bešeňová is located at the boundary of the Liptov Basin (in Slovak–Liptovská kotlina) and the Choč Mountains ca. 12 km west of Liptovský Mikuláš. An aquapark complex with thermal water is located in the centre of the village. An active travertine cascade is being developed ca. 400 m north of the village centre, whereas several outcrops of inactive travertines are located on the slopes above the cascade, mostly in small, inactive quarries (Fig. 16).

The Liptov Basin is one of the intramontane basins in the Central Western Carpathians. It is aligned W–E and located between the Alpine-type mountain chains – the Tatras and the Choč Mts on the north and the Low Tatras on the south (in Slovak – Nízke Tatry). The basin is filled with the Central Carpathian Palaeogene 100 to 2200 m thick (Remšík *et al.*, 2005). These rocks are the regional confining bed whereas underlying carbonates are regarded as an aquifer (Hynie, 1963).

The travertines in the vicinity of Bešeňová are situated on a fault stretching W–E, parallel to the Choč–Tatra Fault which is a main tectonic line bordering the Liptov Basin on the north (Gross *et al.*, 1979; Gross, 1980). Additionally, a W–E stretching fault intersects a meridional fault (Bešeňová Fault – Fendek *et al.*, 2015) bordering from the west the so-called Bešeňová elevation, that is a

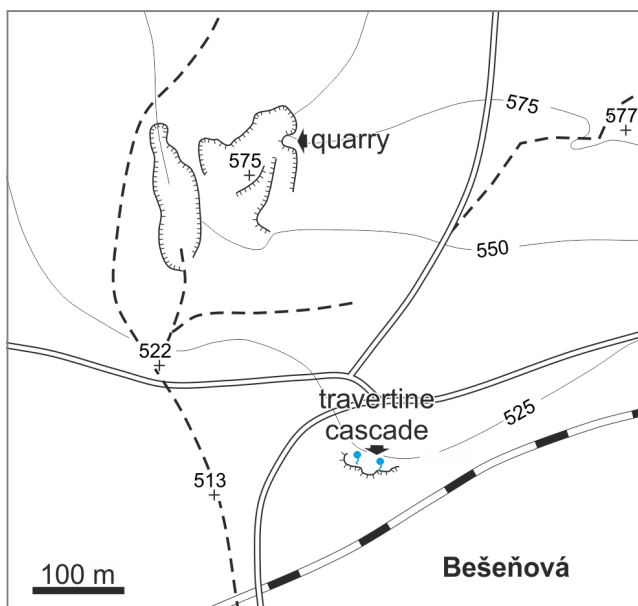


Fig. 16. Location of travertine sites (stops 7.1 and 7.2) presented in Bešeňová.

transverse horst within the Liptov Basin. The faults cut relatively impermeable rocks of the Central Carpathian Palaeogene creating flow path for water ascending from the Mesozoic carbonate aquifer. The aquifer in question is probably a complex hydrological structure whose recharge is supposed to be on the northern slopes of the Low Tatras (Fričovský *et al.*, 2015). The piezometric level of water in the borehole located in the village of Bešeňová is ca. 190 m above the ground surface (Remšík, 2005).

Stop B7.7.1 Bešeňová – Recent travertine cascade

(49°06'14"N, 19°26'09"E)

Leaders: Michał Gradziński, Pavel Bella,

Maria Jolanta Chmiel¹,

Joanna Czerwik-Marcinkowska, Marek Duliński,

Jacek Motyka, Teresa Mrozińska

¹Department of Microbiology, University of Agriculture, Kraków, Poland

The almost 10 m high cascade is a well known tourist attraction. It is developed on the lower part of slopes north of the Bešeňová village (Fig. 16). The cascade is protected as a natural monument. It is fed by water issuing in a series of small springs, some of which are artificially widened. The springs are distant from a few to a dozen metres from the cascade crest. The water is of the Ca–Mg–HCO₃–SO₄ type. It has almost constant temperature throughout the year (between 14.2 °C and 15.6 °C). The water is charged with CO₂ whose pressure at the outflow reaches 1.05 atm. Its mineralization is ca. 3.6 g/L. Mean Ca content is 829 mg/L, HCO₃ content is 3804 mg/L, whereas Fe concentration reaches 0.055 mg/L. The above features suggest that the water is of deep circulation, as is confirmed by the lack of tritium.

Water from the springs flows almost horizontally southward by a system of 'self-built canals'; it reaches the crest of the travertine cascade and forms a thin film seeping down the cascade face. The face is steep, in some segments vertical and covered with microdams and micropools. Precipitation of calcium carbonate is reflected in the chemical evolution of water along its flow path below the lower pool. Concentrations of CO₂ and Ca decrease, whereas pH and saturation index with respect to calcite increase downward.

Travertine formed in the head part of the flow path is deep-orange to intense red in colour, whereas that precipitated in the cascade face is whitish-grey. The former one contains a substantial amount of iron oxides (up to 45 wt%). It is composed of calcite, goethite and amorphous iron oxyhydroxides. The iron compounds precipitate within a slimy biofilm built of filamentous cyanobacteria (mostly different species of *Phormidium*) and diatoms (e.g., *Achnanthes lanceolata*, *A. minutissima*, *Cymbella minuta*, *C. laevis*, *Navicula menisculus*). Iron-bacteria have been also detected (*Metallogenium* sp., *Thiobacillus* sp.).

Travertine precipitated on the cascade face is of crystalline crust type. The spatial arrangement of laminae mirrors the surface of the cascade face, that is its microdam and micropool relief. The exploratory study shows chemical evolution of water along its flow path from the spring to the foot of the cascade as related to CO₂ degassing and precipitation of calcite and iron minerals. Isotopic study of water and modern travertine suggests that the travertine grows in conditions out of isotopic equilibrium.

Stop B7.7.2 Bešeňová, inactive quarry – Pleistocene travertine formed in shallow pond

(49°06'25" N, 19°26'04" E)

Leaders: Michał Gradziński, Marek Duliński, Jacek Grabowski, Helena Hercman, Peter Holúbek¹, Marianna Kováčova, Katarzyna Sobień

¹Slovak Museum of Nature Protection and Speleology, Liptovský Mikuláš, Slovakia

A series of small, abandoned quarries extends above the cascade, on the hill slope called Skala (Fig. 16). They are known as Báňa. The quarries were in use until the second half of the last century. The travertine, with characteristic yellow to pale orange colour, was used as polished building stone for elevations and floors. It was exported to several European countries, including Poland, and to the USA (Pivko, 1999). The quarries are now abandoned and densely vegetated. The clearly visible rock crag crowned with a cross is practically the only available outcrop. The crag is located in the north-eastern part of the quarries. It is protected as a nature monument.

The rock crag is 9 m high. It is built of layered travertine dipping slightly to the west. The crag walls are weathered. Huge travertine blocks partly dismembered and slightly tilted are present in the southern and western parts of the crag. They are separated from the crag by opened and partly karstified fissures.

Laminated micritic travertine is the dominant lithotype in the crag. Calcite rafts and pisoids occur subordinately (Fig. 17). Vertically oriented coated bubbles form individual layers, 1 to 4 cm thick. Such layers are clearly visible, especially in the top part of the crag. Intraclast breccia layers up to 10 cm thick are also present. They are more prone to weathering and they are marked by concave zones in the crag faces. Deformation structures of brittle and ductile type are discernible on the south facing wall of the crag. They are interpreted as a result of a seismic shock in consolidated and unconsolidated travertine, respectively. In spite of its colour, travertine contains only up to 1 wt % of Fe₂O₃.

The travertine lacks faunal remains and plant imprints. Pollen assemblage points at its origin under warm, interglacial climate. U-series dating suggests that it is older than 350 ka but younger than 1.2 ma. Exploratory palaeomagnetic study proves that it is normally magnetized, which implies the age younger than 780 ka.

Facies of the travertine in the rock crag seem to have formed in a shallow pond, or ponds fed with highly mineralized water charged with CO₂ of deep geogenic origin, as evidenced by the δ¹³C values of travertine, ranging from +4.6 to +9.3 ‰ vs V-PDB (Fig. 12; see Gradziński *et al.*, 2008b). The water outflowed along the faults from the Mesozoic carbonates underlying the Central Carpathian

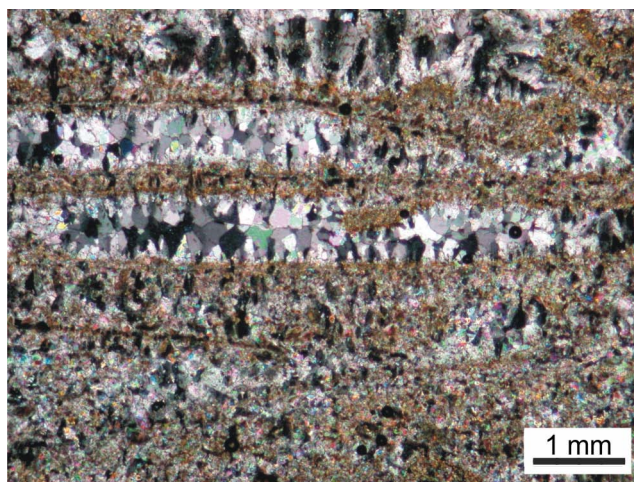


Fig. 17. Calcite rafts cemented with sparry calcite (thin-section), Bešeňová abandoned quarry (stop 7.2).

Palaeogene rocks. Thus, from general genetic point of view, the travertine is an analogue of the modern cascade (Stop B7.7.1). On the other hand, it was laid down in different environmental conditions.

Breccias and crystalline crusts built of phreatic calcite spar are visible on the walls of the rock crag. They filled the fissures which cut the travertine buildup and were exposed during exploitation of travertine. The fissures were also filled with loose deposit, orange, red or pale brown in colour. It comprised rich assemblage of interglacial molluscs (Vaškovský and Ložek, 1972; Vaškovský, 1980) and bones of mammals living in forest and steppe environments (Sabol, 2003). Among others, a skull of a rare bear species (*Ursus taubachensis*) was found. Calcite rafts cemented to a rocky wall are exposed in a small quarry located to the west (Kostecka, 1992). They originated within open fissures during the last glacial.

Stop B7.8 Demänová Cave System – Multi-storey cave system richly decorated with speleothems

(Mramorové rečisko entrance: 49°01'59" N, 19°34'57" E)

Leaders: Pavel Bella, Michał Gradziński, Dagmar Haviarová¹, Helena Hercman, Peter Holúbek, Jacek Motyka

¹Slovak Caves Administration, Liptovský Mikuláš, Slovakia

The Demänová Cave System (DCS; in Slovak – Demänovský jaskynný systém) is situated on the eastern side of the Demänová Valley (in Slovak – Demänovská dolina; Fig. 18; Droppa, 1957). The total length of the DCS exceeds 40 km and its vertical extent equals 202 m (Bella *et al.*, 2014; authors' unpublished data). The system includes ten caves, among which Demänová Cave of Liberty is the longest (in Slovak – Demänovská jaskyňa slobody) and this cave will be visited. The cave lies within a national park, it is protected and its entrances are gated. Some cave parts are accessible for tourists all year round (see <http://www.ssj.sk/en/jaskyna/4-demanovska-cave-of-liberty>). Three stops, namely B7.8.2–B7.8.4 are located on a tourist trail within the cave.

The cave system is developed within Anisian limestones and dolomites of the Gutenstein type. These limestones belong to the allochthonous Krížna Nappe

that makes the northern sedimentary cover of the Low Tatras crystalline core composed of granitoids (Fig. 19; Droppa, 1957; Bella *et al.*, 2014). The upper part of the Demänová Valley was glaciated at least twice during the Middle Pleistocene. The DCS, however, occurs in a narrow canyon located downstream of the glaciated part of the valley and below the preserved till deposits (Droppa, 1972). The DCS originated by corrosion and

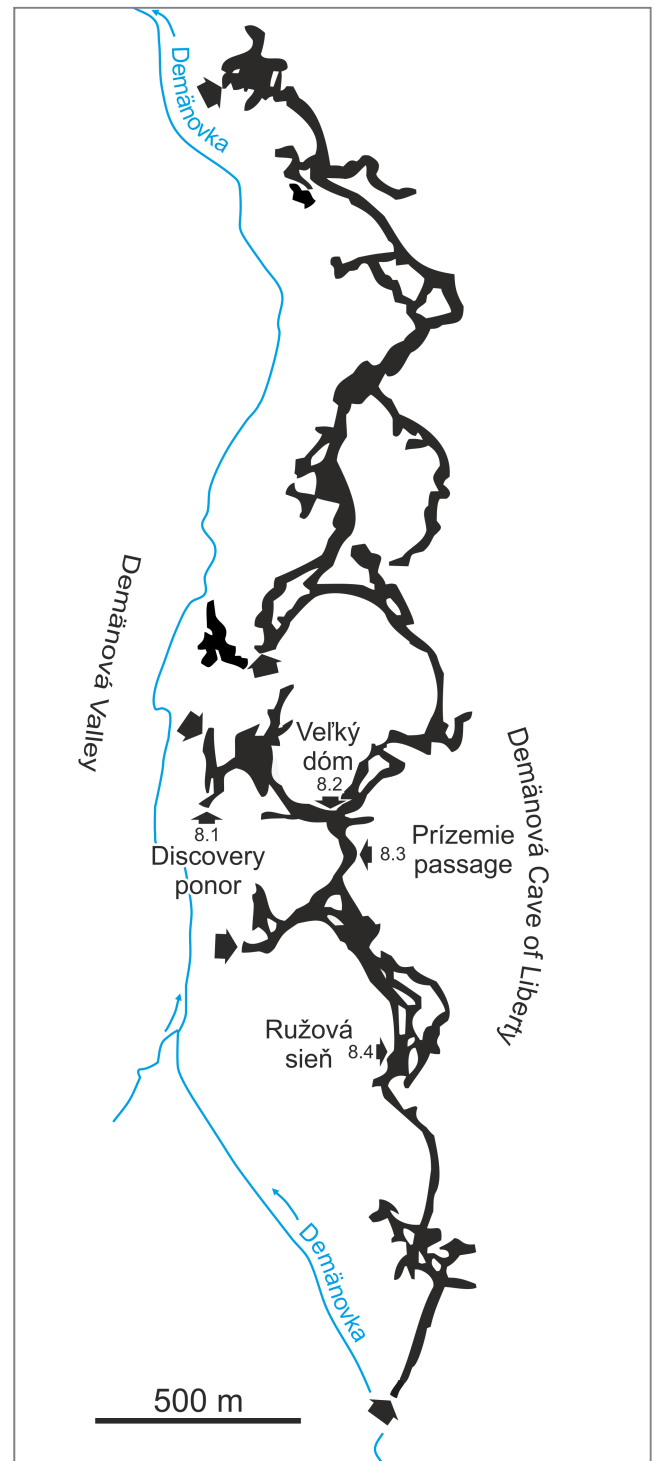


Fig. 18. Map of the Demänová Cave System (after Kučera *et al.*, 1981, simplified), big arrows indicate cave entrances, small arrows indicate stops.

erosion of allochthonous waters. Now the Demänovka stream enters the karst area in Lúčky at the altitude of 950 m, where it sinks underground partly or entirely, depending upon its discharge (Fig. 19).

Droppa (1957) distinguished nine cave levels in the DCS (Fig. 20). In subsequent papers he correlated these levels with the fluvial terraces of the Demänovka stream, and with those of the Váh River and its tributaries (e.g., Droppa, 1966). He assigned individual cave levels to successive glacial stages, using the classical Alpine morphostratigraphic scheme. More recent, detailed studies indicate, however, that the origin of this system was more complicated (e.g., Hochmuth, 1993; Bella et al., 2014). Bella (1993) assigned the individual DCS cave levels distinguished by Droppa (1957) to the ideal watertable caves or to the mixture of phreatic and watertable levelled caves (*sensu* Ford and Ewers, 1978).

Dating of age boundaries in this system by independent physical methods is of crucial importance for further discussion. Isotopic dating of speleothems is the most appropriate method in this respect. Based of the dating

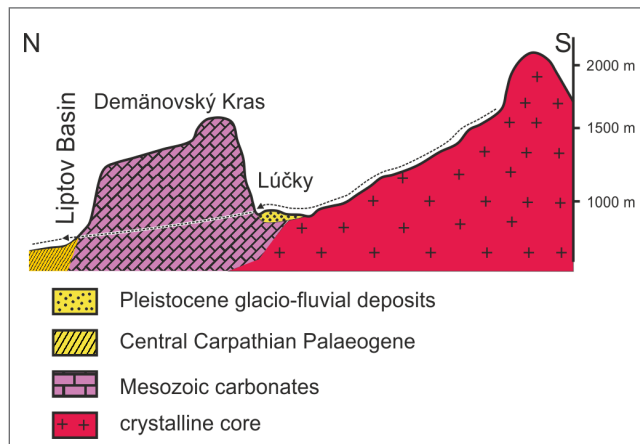


Fig. 19. Simplified geological location of the Demänovka Cave System (after Bella, 1994).

results, one can distinguish in the DCS several generations of speleothems which developed chiefly in the warm periods of Pleistocene and in Holocene (Hercman *et al.*, 1997; Hercman, 2000; Hercman and Pawlak, 2012). It is justified to conclude that the age of the oldest speleothems occurring in a particular cave level of the DCS is the minimum age of dewatering of this level (Hercman *et al.*, 1997).

Stop B7.8.1 Discovery Ponor of Demänová Cave of Liberty

(49°0'3" N, 19°35' E)

The ponor (swallow hole) is situated on the right (eastern) side of the valley, at the altitude of 805 m, close to a tourist trail running along the valley. It is the lowermost ponor of the Demänovka river. Demänová Cave of Liberty (DCL) was discovered by A. Král with the help of A. Mišura and other surveyors through this sinkhole in 1921. The discovery of DCL featured a definite impulse for the development of speleology in Slovakia. A part of the cave from the Mramorové rečisko passage through the Veľký dóm chamber has been opened to the public since 1924 by an old entrance which is situated 8 metres above the Discovery Ponor.

Stop B7.8.2 Demänová Cave of Liberty – Veľký dóm

Veľký dóm is one of the biggest chambers in DCL. Its height reaches 41 m, whereas its lateral extent is 75 m x 45 m (Droppa, 1957). It was formed at the intersection of two fissures striking 300° and 70°. The chamber floor is littered with scree composed of individual blocks up to a few metres across. This material originated from

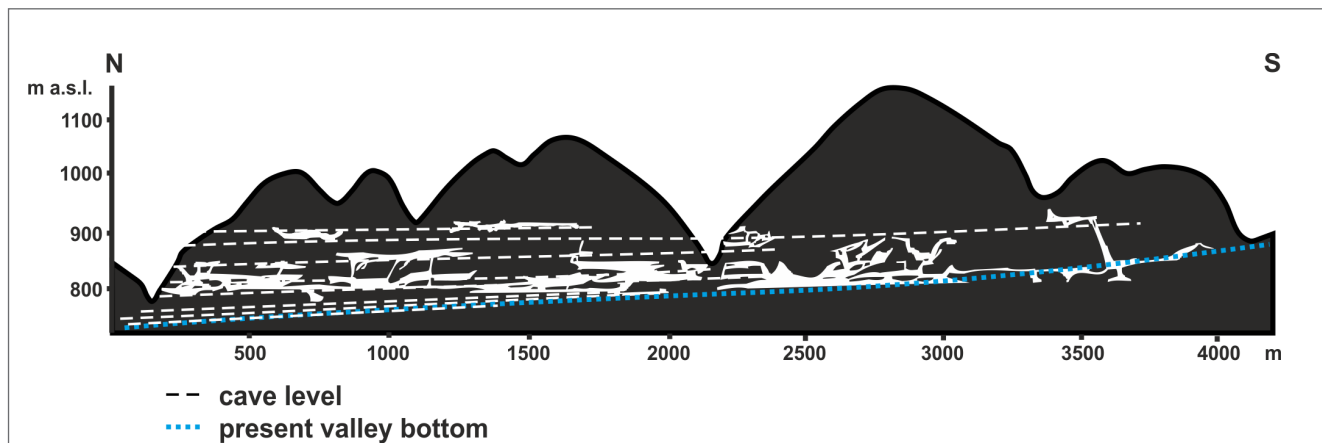


Fig. 20. Longitudinal cross-section of the Demänovka Cave System (after Droppa, 1966, simplified).

a collapse or collapses of the chamber ceiling. The underground segment of the Demänovka river flows below the blocks. The scree is overgrown with stalagmites, some of them up to 1 m high (Droppa, 1957). Dating of the stalagmites revealed their Holocene age (authors' unpublished data), which suggests the age of the youngest collapse event in the chamber. North-east wall of the chamber is covered with extensive active moonmilk flowstone.

Stop B7.8.3 Demänová Cave of Liberty – Prízemie passage

The studied sediment section is located at the level of the active underground course of the Demänovka river. The section is more than 4 m thick (Fig. 21). It is composed of clastic deposits divided by several flowstone floors. The oldest visible deposits consist of gravels covered by flowstones and stalagmites, the oldest of which are dated at 236 ± 10 ka. Some parts of this flowstone are in the actual river bed. Higher flowstone floors that occur within clastic deposits have been dated to the last glacial period and Holocene. The ages of these flowstones prove that the lowermost level of DCL (*sensu* Droppa, 1966) is, at least in this part of the cave, considerably older than the Holocene.

Stop B7.8.4 Demänová Cave of Liberty – Ružová sieň

After walking along several passages and chambers, the trip reaches Ružová sieň, which represents one of the most beautifully decorated parts of DCL. There are several stalagmites and limestone pools constantly filled with water. The hydrochemical study reveals that the water mineralization (TDS) reaches 300 mg/L, pH varies between 7.8 and 8.5. The water represents the

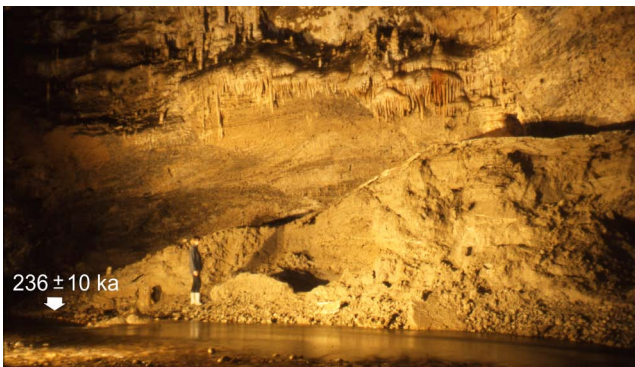


Fig. 21. General view of the Prízemie passage, the age and location of the oldest flowstone are indicated.

Mg–Ca–HCO₃ type. Elevated concentrations of such ions as Al, Cu, Mo, Ni, Pb and Zn, as compared to other underground pools in DCL (see Motyka *et al.*, 2005), most probably results from coins dropped into the studied limestone pool. Saturation indices imply that water is in equilibrium or, sometimes, saturated with respect to calcite and aragonite. This suggests that speleothems constantly grow in this part of the cave, which is in line with their fresh appearance.

Stop B7.9 Lúčky – Travertine fed by a mixture of deep- and shallow-circulating waters

Leaders: Michał Gradziński, Pavel Bella

The village of Lúčky is located on the boundary of the Liptov Basin and the Choč Mountains ca. 15 km WNW of Liptovský Mikuláš (Fig. 22). It is situated straight on the major fault bordering the Liptov Basin from the north and separating it from the Choč Mts and the Tatras (Gross, 1980). Additionally, this fault intersects a meridional one. The Teplianka Valley in the Choč Mts, north of the village, is developed along the latter fault.

The Teplianka stream drains the Choč Mts and flows southward to the Váh river. In its upper course it flows in a narrow, deeply entrenched valley, which is widened near its mouth to the Liptov Basin. The water of the Teplianka stream derives from shallow circulation and its recharge zone is located in the Choč Mts, at altitudes up to 1611 m. It is weakly mineralized (TDS = ca. 320 mg/L) and represents the Ca–Mg–HCO₃ or Ca–HCO₃ types. In the area of the Lúčky spa, this water forms a mixture with artesian water of deep circulation (Franko and Hanzel, 1980). Actively growing travertine is fed by a mixture of both types of water.

Vaškovský and Ložek (1972) recognized three generations of travertines, besides the actively growing ones, near the village of Lúčky. The oldest one crops out in small, abandoned quarries in the south-east outskirts of the village. The second generation provides the most spectacular outcrops. It forms a distinct escarpment and a terrace on the western side of the valley, above the old part of the village and below the spa. The accessible outcrops are artificial – road cuts and abandoned quarries. One of them is presented during the trip (Stop

B7.9.3). The youngest, third generation, is of Holocene age. It crops out in a terrace riser over an artificial lake below the village church. There is a picturesque waterfall, rising up to 15 m over the lake. Modern travertine is being precipitated vigorously on the waterfall (Stop B7.9.2).

Stop B7.9.1 Lúčky spa – Outflow of deep-circulating water

(49°08'05"N, 19°24'14"E)

**Leaders: Michał Gradziński, Pavel Bella,
Marek Duliński, Jacek Motyka**

The Lúčky spa (in Slovak – Lúčky kúpele) lies at the mouth of the Teplianka Valley. The stop is located close to the road, within the spa buildings at the capture of water from borehole Valentína (Fig. 22).

The mineral springs at Lúčky were noticed in a manuscript dated at 1712 and later, in 1736, by Slovak naturalist Matej Bel (Moravčík, 2012). The water has been used for medical purposes since 1761. The natural springs are located at the altitude of 610–620 m. At present, they are captured for the spa which additionally takes water from boreholes. The recharge area may be located in the western part of the Tatras, in the Low Tatras and in the Choč Mts (Hynie, 1963; Franko and Hanzel, 1980; Franko, 2002; Fendek *et al.*, 2015).

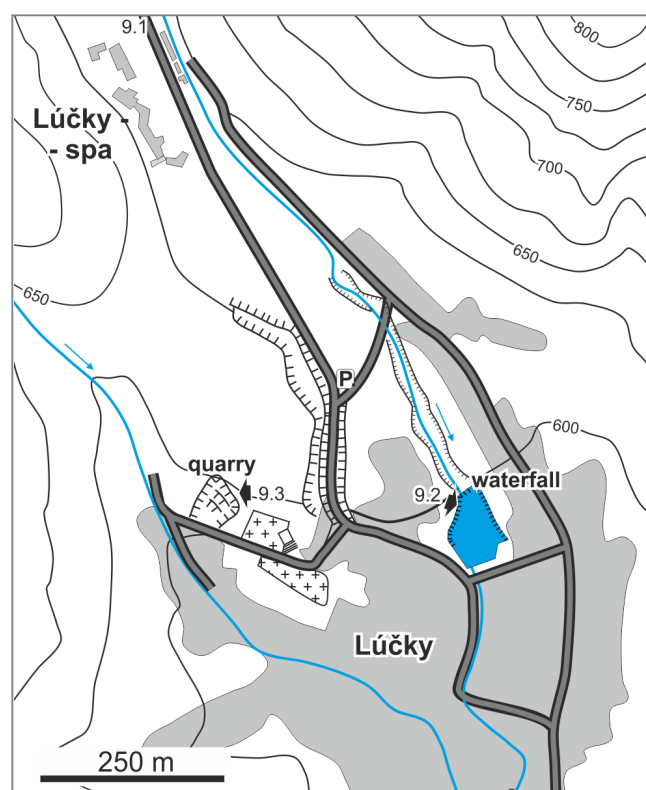


Fig. 22. Location of sites presented in Lúčky (stops 9.1–9.3).

The water from the spa feeds the Teplianka stream.

Mineral water (TDS = ca. 2.6 g/L, temperature at the outflow ca. 30 °C) is exploited from a borehole. It represents the Ca–Mg–HCO₃–SO₄ type under a pressure of ca. 0.5 atm. It represents the Ca–Mg–HCO₃–SO₄ type. Carbon dioxide present in water is probably of mantle origin. Its δ¹³C is close to –6.0 ‰ vs V-PDB. The water does not contain tritium. Stable isotope composition δ¹⁸O and δ²H of this water is close to –10.8 ‰ and –74 ‰, respectively. Thus, it is only slightly lower than that observed for present infiltration waters in this area.

Stop B7.9.2 Lúčky waterfall – Environmental control on deposition of modern travertine

(49°07'47"N, 19°24'12"E)

**Leaders: Michał Gradziński, Pavel Bella,
Joanna Czerwik-Marcinkowska, Marek Duliński,
Peter Holúbek, Jacek Motyka, Teresa Mrozińska**

Waterfall is located in the centre of the village, ca. 100 m south-east of the church (Fig. 22). Waterfall and its vicinity are protected as a nature monument.

Travertine is being precipitated along the riverbed of the Teplianka stream, especially on the waterfall. Upstream of the waterfall the stream flows swiftly down; the water spills over some artificially created dams. The water runs on the terrace tread which is built of detritic deposits of Teplianka and travertines. The thickness of these deposits reaches 30 m, and locally rises to 45 m (Mitter, 1979). The waterfall is located in the northernmost point of a funnel-shaped, 120 m long gully widening to the south. Its steep walls are built of poorly cemented, bedded travertine. Palaeobotanical study by Němejč (1928) indicates its Holocene (Atlantic) age. Exploratory ¹⁴C dating gives 7845 ± 45 years BP. However, this date must be treated with caution due to the so-called ‘dead carbon effect’.

The waterfall main face is exposed to the south. Its height is 12 m. It is engulfed by two concave formations of vertical drops intervened by small horizontal shelves. Presently, the eastern formation is almost dry after construction of an artificial dam above the waterfall, but water distribution on the waterfall changes also naturally. The waterfall face is formed by several overlapping travertine curtains. Three small caves occur behind some curtains. The biggest of them is 10 m long. They represent

an uncommon type of caves, created due to progradation of a travertine (or tufa) cascade.

The waterfall is supplied by the stream whose flow is composed of at least two components. One represents the water of deep circulation, which is characterized in the previous stop. This water is mixed above the waterfall with typical shallow water drained by the stream, transporting dissolved biogenic CO₂, derived from decomposition of organic matter or root respiration. As a result of mixing, the water feeding the cascade is dominated by carbon of deep origin (ca. 70%). This influences the δ¹³C of total dissolved inorganic carbon, which varies between -3.2‰ and +0.2‰ vs V-PDB. The water is oversaturated with respect to calcite. Its saturation index with respect to calcite ranges from 0.67 to 1.27, whereas TDS varies from 507 mg/L to 1023 mg/L (Gradziński, 2010).

Travertine is being precipitated in the streambed over the waterfall, on the waterfall and below it in an artificial channel which leads water through the village. The growing travertine displays great variation of lithotypes (Gradziński, 2010). Porous travertine with encrustation on filamentous cyanobacteria and algae and compact crystalline crust are the most common ones. The former grows in fast-flow settings (see Pedley, 2000; Pedley and Rogerson, 2010). Its formation results from relatively high supersaturation. The vigorous growth of elongated organic filaments seems to be forced by their encrustation with calcium carbonate (Kano *et al.*, 2003; Gradziński, 2010). Algae of genus *Vaucheria* and cyanobacteria of genus *Phormidium* were identified. The former tend to inhabit well irradiated locations and dominate in spring season, whereas the latter are more tolerant to low light intensity. Several taxa of diatoms, other cyanobacteria and algae also occur at the waterfall. Crystalline crust is developed in fast-flow settings under conditions of high supersaturation (see Pedley, 2000). Its formation results from fast abiogenic crystallization of calcite, which impedes colonization and growth of cyanobacteria and algae. In spite of specific character of carbonate system, which is dominated by the carbon of deep origin, the precipitated travertine records markedly even subtle hydrological changes in the catchment. The episodes of fast growth reflect cessation of stream-water supply from the catchment. Conversely, laminae abounding in detritic components record increase in surface runoff (e.g., snow-melt episodes; Gradziński, 2010).

The mean growth rate of travertine at the Lúčky waterfall, studied between 2002 and 2003, was 4.994 mg·cm⁻²·day⁻¹ (Gradziński, 2010). It definitely exceeds the rate of growth of tufa fed with water of shallow-circulation but it is lower than the growth rate of thermal travertine, which can reach 30.9 mg·cm⁻²·day⁻¹ (Pentecost and Coletta, 2007).

The values of δ¹³C and δ¹⁸O of the travertine vary from +0.6‰ to +3.3‰ and from -10.9‰ to -8.6‰ vs V-PDB, respectively. The comparison of the above values with the isotopic parameters of feeding water implies that oxygen isotopes are not kinetically fractionated during calcite crystallization, whereas carbon isotopes of travertine are affected by kinetic fractionation, probably in CO₂ outgassing process.

Stop B7.9.3 Lúčky, inactive quarry – Facies of Pleistocene travertine fed by mixed water

(49°07'48" N, 19°23'54" E)

Leaders: Michał Gradziński, Pavel Bella, Marek Duliński, Helena Hercman, Peter Holúbek, Ewa Stworzewicz¹

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Yellow bedded travertine crops out in an inactive quarry near the village cemetery on the north-west outskirts of the village (Fig. 22). The quarry is protected as a nature monument. The beds dip towards the south-east at an angle of 15° to 65° which reflects the preexisting relief. Travertines are mostly represented by two facies: (i) phytoclastic travertine which abounds in leaf imprints, twig and stem empty moulds, and (ii) cyanobacterial and algal stromatolites. Moss tufa occurs subordinately.

In the western part of the quarry, where the oldest rocks are exposed, phytoclastic travertine predominates. The empty moulds of twigs and stems are horizontally or subhorizontally oriented. Some of them reach diameter of 0.5 m and lengths up to 3.5 m (Gradziński, 2008). Leaf and tree-needle imprints are common, cone imprints occur as well. Grass blades imprints are in life position. They are cemented by sparry calcite, which most probably encrusted algae and cyanobacteria.

Stromatolites crop out in the central and eastern part of the quarry. They were constructed by filamentous cyanobacteria and algae (Fig. 23). In the uppermost beds, a level with gravel of Triassic carbonates, partly covered with stromatolites, has been found.

Snail shells occur within both facies of travertine. They are numerous in some places. They represent land and freshwater snails (Vaškovský and Ložek, 1972).

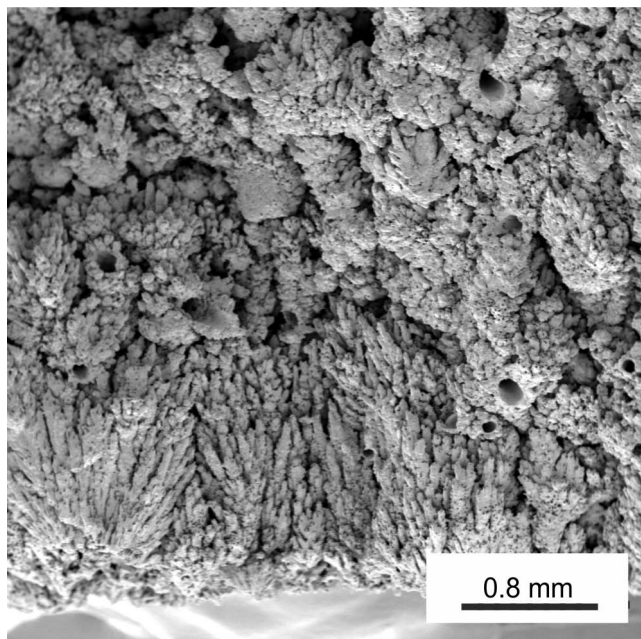


Fig. 23. Stromatolites composed of cyanobacteria (radiating fans in the lower part), which alternate with algae (empty moulds in the upper part), SEM image.

The age of the lower part of the section, determined by U-series method, is equal to t_{ka} , whereas the age of upper part is equal to t_{ka} . This is in agreement with the former opinions by Němejc (1928) and Vaškovský and Ložek (1972) based on palaeobotanical and malacological data, respectively.

Facies of the travertine which crops out in the quarry bear a strong resemblance to those of calcareous tufa which is fed with meteoric water of shallow circulation (see Pedley, 2009; Vázquez-Urbez *et al.*, 2012). However, the values of $\delta^{13}C$ of travertine discussed fall between -0.4‰ and $+4.5\text{‰}$ vs V-PDB (Fig. 12). Such carbon isotopic composition is different from composition typical for calcareous tufa. Conversely, it clearly indicates that the travertine was intimately associated with geogenic CO_2 . Thus, one can presume that it can be regarded, to a great extent, as an analogue of recent travertines observed at the Lúčky waterfall.

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