Annales Societatis Geologorum Poloniae (2016), vol. 86: 341–350. http://dx.doi.org/10.14241/asgp.2016.016

RADIOCARBON DATING OF FOSSIL BATS FROM DOBŠINA ICE CAVE (SLOVAKIA) AND POTENTIAL PALAEOCLIMATIC IMPLICATIONS

Michał GRADZIŃSKI[1], Helena HERCMAN[2], Andrea PERESWIET-SOLTAN[3, 4], Ján ZELINKA[5] & Magdalena JELONEK[1]

[1] *Institute of Geological Sciences, Jagiellonian University, Oleandry 2a, 30-063 Kraków, Poland; e-mail: michal.gradzinski@uj.eu.pl (MG), magdalena.lata@hotmail.com (MJ)*
[2] *Institute of Geological Sciences, Polish Academy of Sciences, Twarda 51-55, 00-818 Warsaw, Poland; e-mail:hhercman@twarda.pan.pl*
[3] *Institute of Systematics and Evolution of Animals, Polish Academy of Sciences, Sławkowska 17, 31-016 Kraków, Poland; e-mail: pereswiet_soltan@yahoo.it*
[4] *Club Speleologico Proteo Vicenza, Viale Riviera Berica, 631 - 36100 Vicenza, Italy*
[5] *Slovak Caves Administration, Hodžova 11, 031 01 Liptovský Mikuláš, Slovakia; e-mail: jan.zelinka@ssj.sk*

Gradziński, M., Hercman, H., Pereswiet-Soltan, A., Zelinka, J. & Jelonek, M., 2016. Radiocarbon dating of fossil bats from Dobšina Ice Cave (Slovakia) and potential palaeoclimatic implications. *Annales Societatis Geologorum Poloniae*, 86: 341–350.

Abstract: Although Dobšina Ice Cave (DIC, Carpathians, Slovakia) is located outside the high-mountain area, it hosts one of the most extensive blocks of perennial subterranean ice, the volume of which is estimated at more than 110,000 m^3. Frozen bat remains were found in the lowermost part of the perennial ice block. They belong to *Myotis blythii* (Tomes) and the *M. mystacinus* morpho-group. The radiocarbon dating of bat soft tissues yielded ages of 1266–1074 cal. yr BP and 1173–969 cal. yr BP. The undetermined bat, found in the same part of the ice section in 2002, was previously dated at 1178–988 cal. yr BP (Clausen *et al.*, 2007). The dates testify that the ice crystallized at the turn of the Dark Ages Cold Period and the Medieval Warm Period. The calculated accumulation rate of cave ice varies between 0.7 cm/year and 1.4 cm/year at that time, and is similar to the present ice accumulation rate in DIC. Constant crystallization of ice during the Medieval Warm Period is hypothesized to reflect dry summer seasons since the supply of relatively warm water in the summer is one of the key factors causing the erosion of cave ice. The uppermost sample was covered with 20.6 m of ice. Between ca 1065 cal. yr BP and the present day, the ice grew faster than between ca 1210 yr BP and ca 1065 yr BP by a factor of 1.3–1.8. This may have resulted from conditions favourable for ice accumulation during the Little Ice Age.

Key words: Little Ice Age, Medieval Warm Period, perennial cave ice, Western Carpathians.

Manuscript received 2 September 2016, accepted 4 November 2016

INTRODUCTION

There has been a growing interest in palaeoclimatic research in recent years since knowledge of past climatic changes can shed substantial light on such changes that have occurred recently and will occur in a near future. Various continental deposits have been extensively studied to reconstruct the past climates and climatic changes. Cave deposits, especially carbonate speleothems, have been widely recognized as reliable climatic proxies (e.g., Fairchild and Baker, 2012 and references therein). As such, they have been discussed in several papers. Conversely, cave ice has attracted less attention; it has been the subject of detailed studies only during the last twenty years (Laursen, 2010). Cave ice is particularly important since it occurs commonly outside the zone of existence of glacial ice. Thus, it can be regarded as a cryospheric archive complementary to glacial ice that is widely recognized and has provided a valuable palaeoenvironmental record. Cave ice is particularly worth studying, since the general tendency to lose of its mass has been noted in several caves worldwide (Kern and Perşoiu, 2013). This tendency has been documented in the caves of Siberia (Trofimova, 2006), the Swiss Jura (Luetscher *et al.*, 2005), the Austrian Alps (Mais and Pavuza, 2000; Behm *et al.*, 2009; Spötl *et al.*, 2014), the Polish Tatra Mountains (Rachlewicz and Szczuciński, 2004; Hercman *et al.*, 2010;

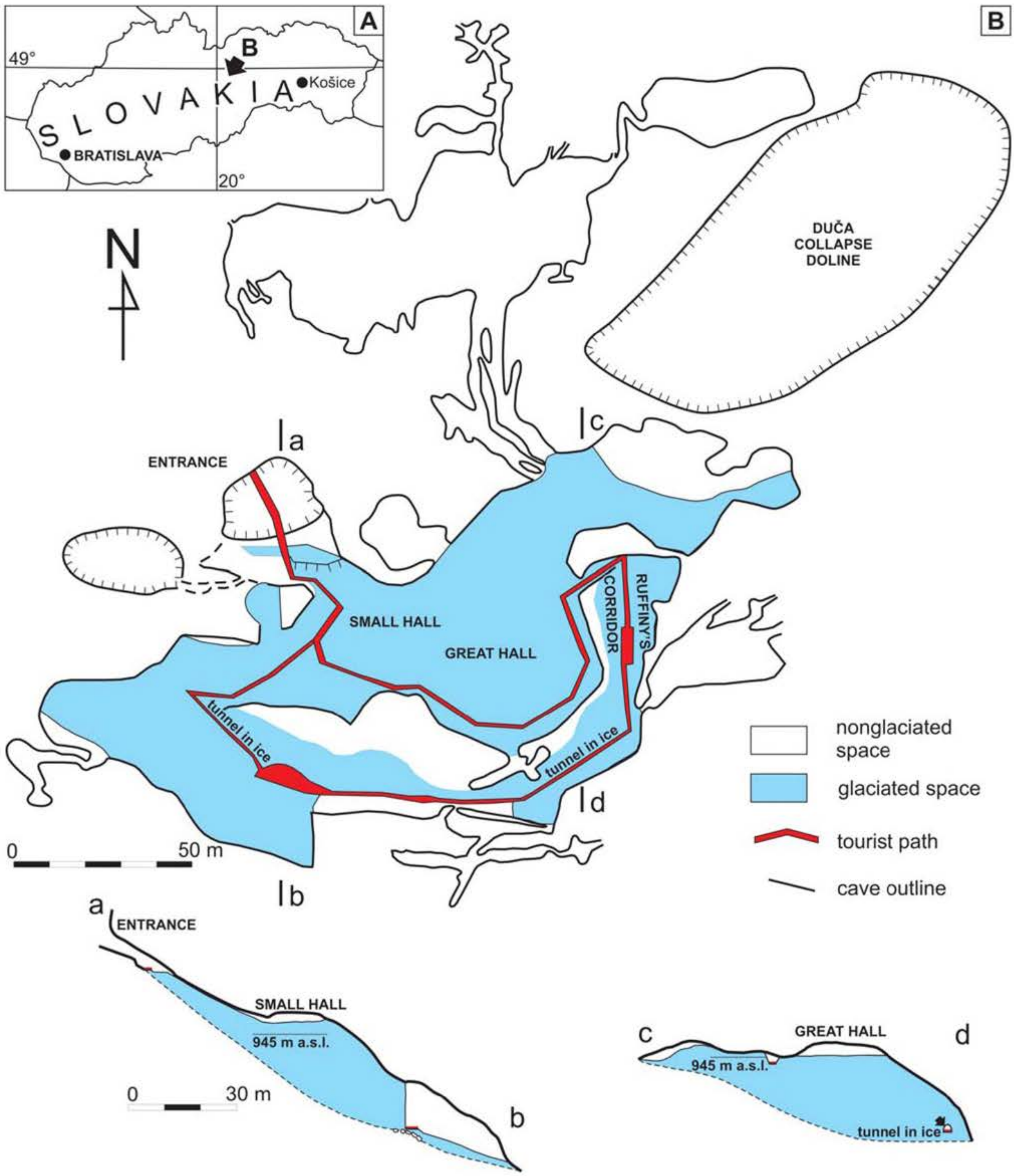

Fig. 1. Location of DIC (**A**), map and cross-sections (**B**) of the cave (after unpublished data by L. Novotný and J. Tulis, simplified), arrow indicates sampling site.

Szukała, 2010) and various regions of the United States (Fuhrmann, 2007; Kern and Thomas, 2014), including the Hawaii (Pflitsch *et al.*, 2016).

The genesis and mass balance of cave ice are strongly dependent upon the climatic context of a given region, cave morphology, as well as the local microclimatic conditions in a cave and in its surroundings (Luetscher *et al.*, 2005; Colucci *et al.*, 2016). Many papers present the results of detailed studies on the behavior of cave ice as a result of the modification of the outside temperature and humidity throughout the year (e.g., Rachlewicz and Szczuciński, 2004; Obleitner and Spötl, 2011; Perşoiu and Pazdur, 2011; Schöner *et al.*, 2011; Yang and Shi, 2015). Thus, the presence of older (subrecent) cave ice indicates the former persistence of particular conditions, favorable for its accumulation.

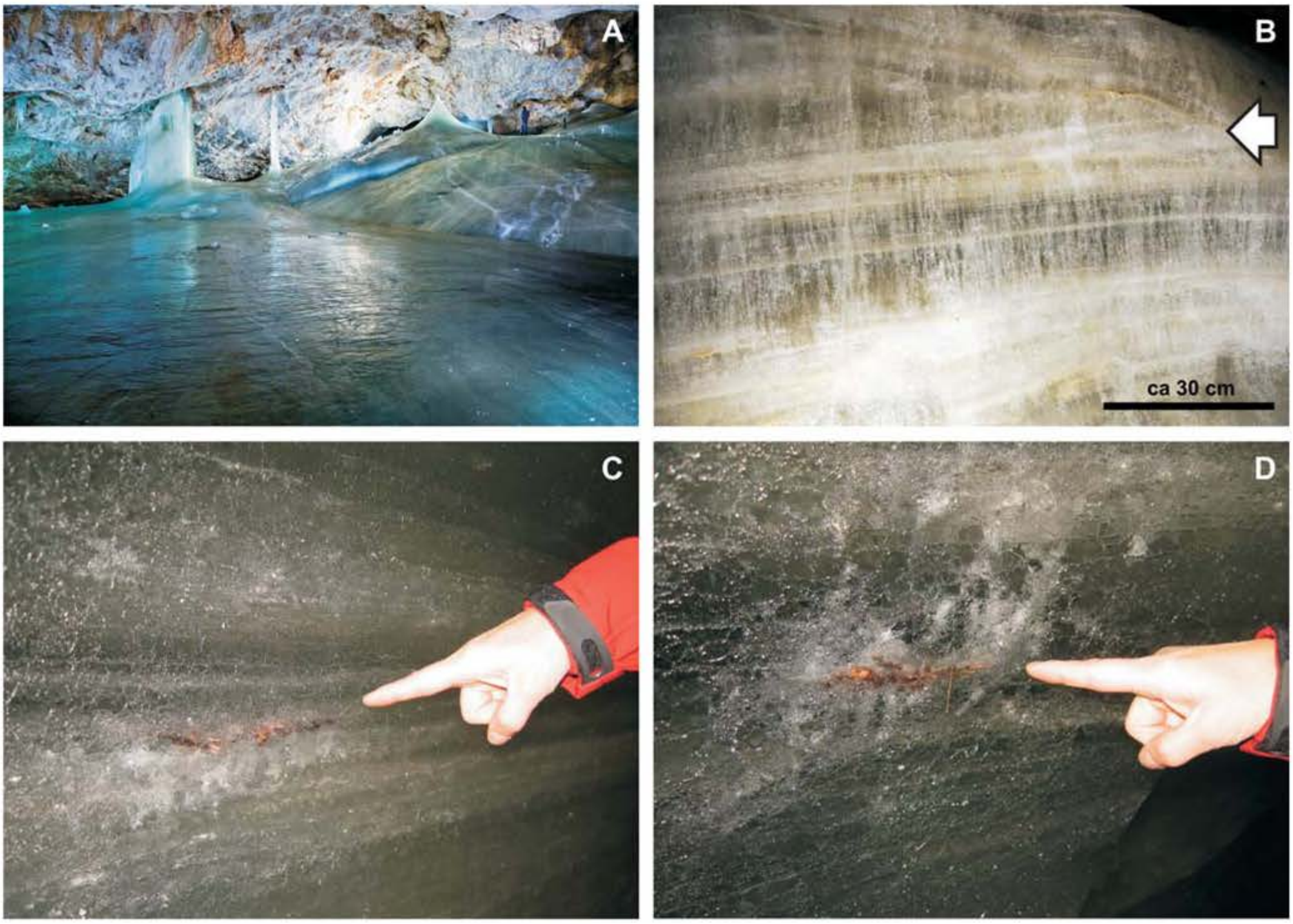

Fig. 2. Cave ice in DIC. **A.** Great Hall with floor ice and ice columns, a caver for scale in the upper right-hand corner; photo by M. Rengevič. **B.** Ice section, transparent layers alternate with opaque layers; thermo-erosion surface is arrowed. **C, D.** Samples NIET-1 and NIET-2, respectively, before extraction from cave ice.

Cave ice is an important record of several palaeoenvironmental data, not only of palaeoclimatic significance (see Kern and Perşoiu, 2013, for review). The hydrogen and oxygen isotopic composition of cave ice bear some information on the isotopic composition of the parent water (Yonge and MacDonald, 1999; Kern *et al.*, 2004, 2011a). The chemical composition of the parent water and the presence of some trace elements can be studied by glaciochemical research (e.g., Citterio *et al.*, 2004; Clausen *et al.*, 2007; Kern *et al.*, 2011b). Pollen entrapped within cave ice allows reconstruction of the type of vegetation cover (Feurdean *et al.*, 2011).

Proper decoding of the palaeoenvironmental information recorded in cave ice requires unravelling its age. Thus, the dating of cave ice is of crucial importance for any palaeoenvironmental studies based on the cave ice archive and several methods of dating have been applied (e.g., Wilson, 1998; Luetscher *et al.*, 2007). Radiocarbon dating of organic debris entrapped within cave ice seems to be one of the most effective methods. Tree trunks and branches, which commonly occur within ice in the entrance zones of caves, have been dated in many caves. Ice in the deeper zones of some caves commonly lacks organic debris, excluding some particulate organic matter, which is not easy to date owing to technical obstacles (May *et al.*, 2011).

The dating of several samples from one ice section allows the construction of age-depth models to estimate the ice accumulation rate. The ice accumulation rate carries important information about former climatic conditions in a given area. Such studies have been carried out in a few caves only, for example, in Scărişoara Ica Cave (Perşoiu and Pazdur, 2011), Hundsalm Eishöhle und Tropfsteinhöhle (Spötl *et al.*, 2014).

This study focuses on the interpretation of the dating of bat remains that were entrapped in cave ice in Dobšina Ice Cave (in Slovak – Dobšinská ľadová jaskyňa, DIC hereafter in this paper). DIC hosts one of the largest blocks of cave ice known worldwide. The dating results provide a basis for estimates of the ice accumulation rate and set the stage for palaeoclimatic studies, arising from the analysis of various proxies recorded by the ice block.

SPELEOLOGICAL SETTING

DIC is situated in northern Slovakia, in the Slovenský Raj mountains (Fig. 1). The cave totals 1,483 m in length, whereas its vertical extent is 112 m (Bella, 2008). The cave was formed in the Neogene as a part of an extensive karst drainage system, hosted by the Middle Triassic (Steinalm

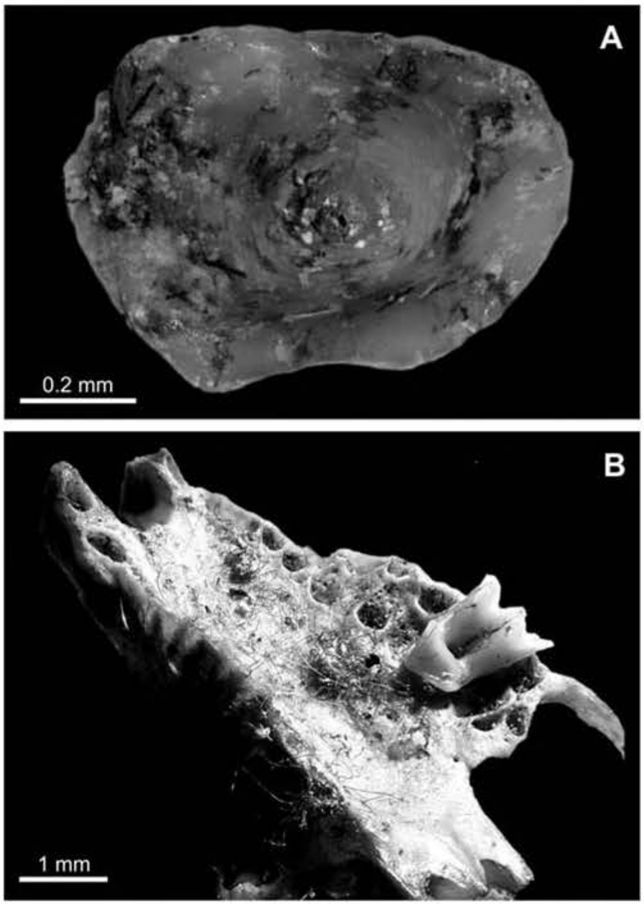

Fig. 3. Bat remains. **A.** *Myotis blythii* – right P_3. **B.** *Myotis mystacinus* morpho-group – rostrum with left M^2.

and Wetterstein type) carbonates of the Silica Nappe (Tulis and Novotný, 1989). The system was dismembered, owing to collapse manifested at the surface as a large doline, called Duča (Fig. 1). Stratenská Cave, with a length of almost 22 km, is another, larger part of the same system. The DIC entrance is located on the north-western slope of Duča hill at an altitude of 969 m (N48°51′48″, E20°18′43″). The cave is located below the entrance; it consists of a huge chamber, the floor of which descends to the south-east from the entrance. The chamber volume is estimated to be over 140,000 m^3 (Zelinka, 2008). The chamber is partly filled with an extensive ice block that reaches the ceiling of the chamber in its southernmost part (Fig. 2A). Ice forms as a result of the freezing of percolation water. It represents a congelation ice type, fed by a thin sheet of water. The ice surface is estimated at more than 9,000 m^2 and its volume at more than 110,000 m^3 (Bella, 2008). The upper surface of the ice block is flat and horizontal in some parts, in others it is inclined towards the south-west, south and south-east (Strug, 2011, ryc. 14). Ice columns and stalagmites up to 10 m high exist on the upper surface of the ice block. A rimaye (called Ruffiny's Corridor) between the cave wall and the ice block is formed in the south-eastern part of the chamber (Fig. 1). A section of layered ice ca 23.5 m thick is exposed there.

Although DIC is characterized by a mixed static-dynamic microclimate, the chamber with the ice block acts as a cold-air trap (Zelinka, 2008). The mean temperature in the glaciated part of the cave varies between –0.4 and –1.0 °C; in the lowermost parts of the cave it is below the freezing point all year round (Bella, 2008; Korzystka *et al.*, 2011, fig. 6). The temperature in the cave strongly depends upon air circulation and exchange. The cold air descends to the lower part of the chamber during the winter seasons. Ice crystallizes mainly during the spring seasons, when infiltration water from thawing snow enters the frozen cave (Korzystka *et al.*, 2011). During the summer, when the temperature of the cave interior slightly increases, the cave ice starts melting. The melting is especially intensive during wet summers, which is connected with the copious amount of relatively warm water infiltrating into the cave (Korzystka *et al.*, 2011). Flowing water and dripping water create various ablation forms on the top of the ice block in DIC under such conditions (Bella, 2007).

MATERIALS AND METHODS

The section studied is located in the south-eastern part of the cave in a tunnel artificially dug in cave ice and on the ice wall of the rimaye corridor called Ruffiny's Corridor. It comprises a vertical range of between ca 922 and 945 m a.s.l. During careful examination of an ice wall, two samples with bone fragments, visible to the naked eye, were extracted from the ice (Fig. 2C, D). They had emerged partly from the ice, owing to sublimation. Samples were carefully extracted by means of a geological hammer and transported to the laboratory, where they were allowed to air-dry. Amorphous-looking remnants of bat soft tissue were separated from the bone fragments, which subsequently were determined after rinsing them with distilled water and examination under a binocular microscope.

The remnants of bat soft tissue were dated in the Poznań Radiocarbon Laboratory using the AMS technique (Goslar *et al.*, 2004). Samples were treated with the standard "acid-base-acid" procedure. For the next step, CO_2 was produced by combusting the sample in closed quartz tubes, together with CuO and Ag wool at 900 °C for over 10 hours. The gas obtained was reduced with hydrogen (H_2), using Fe powder as a catalyst. The mixture of carbon and iron obtained was then pressed into a special aluminium holder (Czernik and Goslar, 2001). The standard samples, the samples not containing ^{14}C (coal) and international modern ^{14}C standard (Oxalic Acid II) were prepared in the same manner. The ^{14}C concentration of the samples was measured, using the "Compact Carbon AMS" spectrometer produced by the National Electrostatics Corporation, USA. The standard sample pretreatment procedure was used (Czernik and Goslar, 2001). Quoted errors are 1 standard deviation and the age is a conventional radiocarbon age. The dates obtained were calibrated using OxCal v. 4.2 (Bronk Ramsey, 2009) with the IntCal13 calibration curve for atmospheric data from Reimer *et al.* (2013). The stratigraphic position of the samples studied in the ice profile was taken into account using the "sequence model" OxCal option. Modelled age distributions from the sequence model were used for the age-depth model construction. The age-depth model (chronology) was constructed, using MOD-AGE software (Herc-

Table 1

Radiocarbon dating, calibration and MOD-AGE results

Sample	Laboratory number	Position below present-day ice surface [m]	Conventional age [yr BP]	Unmodelled [cal. yr BP]			Modelled [cal. yr BP]			MOD-AGE age with 2σ standard deviation range [yr BP]
				from	to	probability [%]	from	to	probability [%]	
Clausen *et al.* (2007)	KIA 23374	20.6	1168 ± 28	1178 1030	1048 988	80.9 14.5	1141	976	95.4	1065^{+95}_{-80}
NIET-2	Poz-53813	21.45	1140 ± 30	1173 1148 1095	1157 1102 969	5.5 11.3 78.6	1175	1011	95.4	1125^{+60}_{-80}
NIET-1	Poz-53812	22.06	1240 ± 30	1266 1163	1170 1074	61.3 34.1	1270	1090	95.4	1210^{+60}_{-120}

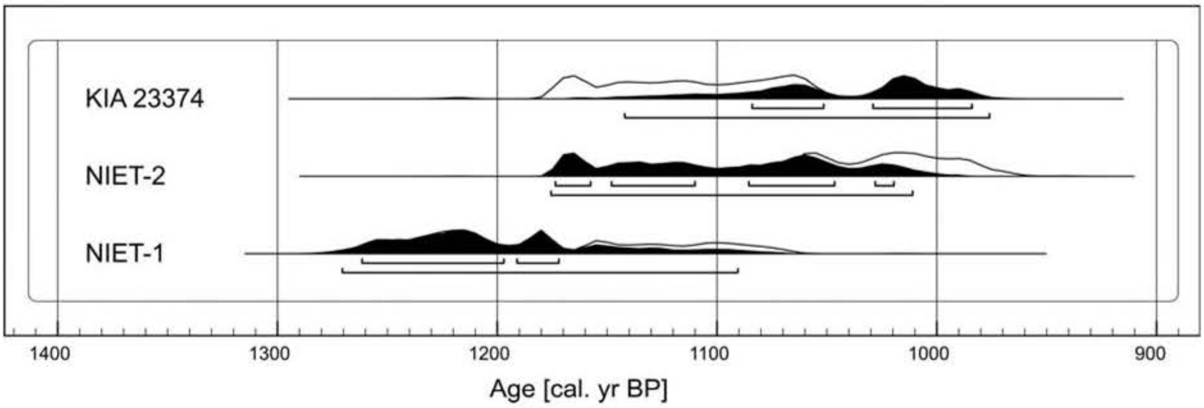

Fig. 4. Results of radiocarbon data calibration using sequence model; black-coloured distributions take into account stratigraphical order of analyzed samples in the section

man and Pawlak, 2012). MOD-AGE takes into account the full modelled age distribution as well as depth error estimation. Depth uncertainties for model construction were assumed as normally distributed with a standard deviation of ± 2 cm. Monte Carlo simulation was used for estimation of the best age-depth relations and its 2σ range. The chronology was estimated as the median of 5,000 possible age-depth models.

RESULTS

The section studied, located in the lowermost part of the cave, totals 23.5 m. The upper part of the section is accessible only with great difficulty for observation; however, some thermo-erosion surfaces are visible in this part (Fig. 2B). The section displays layered ice; transparent layers alternate with opaque, whitish layers. The latter contain bubble inclusions that reduce the ice transparency (Clausen *et al.*, 2007). Opaque layers contain admixtures of fine-grained calcite, too. The calcite also is concentrated in distinct horizons clearly visible, owing to their yellow colour. Angular limestone debris also is embedded within the ice.

Both samples contain remnants of bat soft tissue and some bone material. Long bones are present in both samples. Apart from these, the samples contain determinable fragments. The single tooth right P_3 of *Myotis blythii* (Tomes), along with the rostrum with left M^2 of whiskered bats (*M. mystacinus* morpho-group – Dietz *et al.*, 2007), were determined in sample NIET-1 (Fig. 3). Sample NIET-2 contained the toothless mandible and the rostrum with the right M^1–M^2 of a whiskered bat.

Radiocarbon analyses and calibration results are listed in Table 1 and presented in Fig. 4. The radiocarbon data published previously by Clausen *et al.* (2007) was recalibrated using a new calibration curve and are appended in the Table 1. The age of the older sample NIET-1 (Poz-53812) is 1266–1074 cal. yr BP (AD 684–876). The age of the younger one, NIET-2 (Poz-53813), is 1173–969 cal. yr BP (AD 777–981). Previously published data (KIA 23374) from the sample located 85 cm higher in the section than the samples studied is 1178–988 cal. yr BP (AD 772–962). The age-depth model and calibrated age distributions for the samples studied are presented in Fig. 5. The calibrated ages are presented as 2σ ranges from the sequence model. On the basis of the age-depth relations and assuming continuous ice accumulation for the studied part of the section, it is possible to estimate the average accumulation rate between the dated samples. The accumulation rate between NIET-1 sample and NIET-2 sample is 0.7 cm/year, between NIET-2 and KIA

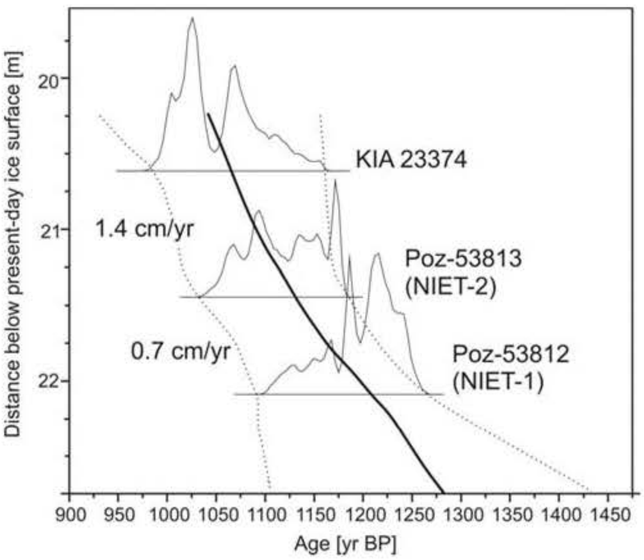

Fig. 5. Age-depth model for lower part of the ice section studied. Thick line – relations of age and depth; dotted line – 2 standard deviation confidence band for estimated relations; radiocarbon age probability distributions located at the sampling depth; mean accumulation rate between NIET-1 (Poz-53812) and NIET-2 (Poz-53813) as well as between samples NIET-2 (Poz-53813) and KIA 23374 are specified.

23374 samples is 1.4 cm/year, whereas between the NIET-1 and KIA 23374 samples it is 1 cm/year (Fig. 5).

DISCUSSION

Meaning and confidence of radiocarbon dates

Organic remains that can be dated by the radiocarbon method are rather scarce in the deeper parts of caves. Although some small quantities of particulate organic matter occur within cave ice even there, the attempts at dating them did not provide fully conclusive results (e.g., May *et al.*, 2011). Easily dated plant debris is common in threshold zones of caves, especially beneath their vertical entrances and is embedded within ice that originated from the diagenesis of snow (e.g., Stoffel *et al.*, 2009; Spötl *et al.*, 2014). Scărişoara Ice Cave can serve as one of the exceptions, since plant debris occurs there in congelation ice, in the deeper part of the cave. However, even there such occurrences result from their location beneath the spacious, vertical entrance (Feurdean *et al.*, 2011; Perşoiu and Pazdur, 2011). Thus, fossil bat remains, including their soft tissue, frozen in layered congelation cave ice at a distance of ca 100 m from the cave entrance, can be regarded as rare and promising objects for dating.

Both samples were found in horizontally layered ice (Fig. 2C, D). The ice that hosted the samples studied does not display any concentration of clastic material. This, in turn, indicates that the samples most likely were not exposed from older ice as a result of ablation events and subsequently refrozen within younger layers of ice.

The interpretation of dating of any organic fossil remains embedded within cave ice must be treated with caution. A lag period is likely to have occurred between the death of a given organism and its being embedded in cave ice (see Spötl *et al.*, 2014), depending on the place of death. In the case of organic debris, especially massive plant fragments introduced from the ground surface, the lag period can be very long. Conversely, bats represent the autochthonous fauna dwelling in caves. Lag periods between a bat's death and its being embedded in cave ice is, more probably, substantially shorter. Thus, the dating results point to the age of the ice with confidence. Some doubts can arise from the co-occurrence of at least two bats in sample NIET-1. However, it could have resulted from their mass mortality during hibernation (Wołoszyn, 1970). The concentration of bat remains by cave dwelling reptiles as a consequence of their predation on the bats could be another explanation. Obuch (2012) drew attention to the remains of the food reserves of the pine marten (*Martes martes* Linnaeus) and also noted that the tawny owl (*Strix aluco* Linnaeus) hunted bats near the cave entrance.

Accumulation history of ice in DIC

The dating results of samples NIET-1 and NIET-2 fully confirmed the result of the previously dated sample KIA 23374. They collectively show that the lower part of the ice complex in DIC accumulated between ca 1210 and 1065 yr BP (ca AD 740 and AD 885). This age is definitely younger than the age suggested by Droppa (1960). He estimated an age of 5000–7000 yr BP or 4133 year BP. The former age was calculated by dividing the maximal ice thickness by the thickness of a single lamina. This concept is based on the assumption that each laminae originates during one year. The latter age was obtained by dividing the maximal ice thickness by a one-year increment detected in the early 1950s, which equals 3 cm/yr. All three dates obtained indicate the formation of ice at the end of the relatively cold period called the Dark Ages Cold Period and the beginning of the subsequent Medieval Warm Period (hereafter MWP), which was dated as the 9th–13th centuries.

The lowermost part of the ice body is older than the oldest sample dated, the age of which is ca 1,210 yr BP (ca AD 740; Table 1, Fig 5). Bearing in mind the results of age modelling by the MOD-AGE program, which takes into account radiocarbon age of the sample and its position in ice section, the oldest ice exposed presently below dated samples crystallized in ca 1280 cal. yr BP (ca AD 670; Fig. 5). However, the above estimation does not clarify when the ice started growing in DIC, since the contact between the rocky bottom and the ice is not visible in this part of the cave (Fig. 1). Moreover, the basal melting phenomenon was documented in the cave (Tulis and Novotný, 2003). Thus, it is not known how much ice melted out from the base of the ice complex. It is very probable that ice crystallized at least during the Dark Ages Cold Period and, plausibly, even earlier.

The accumulation rate between NIET-2 and KIA 23374 was 1.4 cm/yr, whereas during NIET-1 and KIA 23374 it was 1 cm/yr. If the ice grew consistently until the present day at these rates, the uppermost sample KIA 23374 would be covered with ice with a thickness of ca 15.8 m or 11.3 m, respectively. In fact, the ice thickness (up to the present-day surface of the ice block) over sample KIA 23374 is 20.6 m

(Vrana *et al.*, 2007). These collectively indicate that between ca 1065 cal. yr BP (AD 885) and the present day, the ice grew faster than between ca 1210 yr BP (ca AD 740) and ca 1065 yr BP (ca AD 885) by a factor of 1.3–1.8. In fact, this factor must have been even higher, since some ice was melted out, which is recorded as thermo-erosion surfaces (Fig. 2B). This estimate could be explained by analyzing the climatic history of the region. A relatively cold period, called the Little Ice Age (hereafter LIA), set in after the MWP (Jones *et al.*, 2001). The LIA was variously dated to between the 11th and 19th centuries; but persisted roughly between the 15th and 19th centuries. The LIA was characterized in Europe by advances of the Alpine glaciers, a decrease of the tree-line altitude, high lake levels, and changes in tree-ring width (e.g., Haas *et al.*, 1998; Holzhauser *et al.*, 2005; Gąsiorowski and Sienkiewicz, 2010). There is general agreement that this period was drastically colder than the MWP and present-day times, although climatic conditions varied spatially and temporally over Europe at that time (Luterbacher *et al.*, 2016). For example, Popa and Kern (2009) on the basis of tree-ring width postulated substantial differences between the Romanian Carpathians and the Alps in mean summer temperature at the beginning of the 18th century. Feurdean et al. (2015) documented that the LIA was relatively dry in the Carpathians, which may have limited the summer melting of cave ice and, in consequence, may have stimulated a positive mass-balance of cave ice in DIC at that time.

The thermo-erosion surfaces observed in the upper part of the ice section have not been dated. Their position in the ice section indicates formation during the decline of the LIA. One can hypothesize that they originated during a short time period. Dendrochronological data from the nearest area (at the south foot of the Tatra Mountains) document some extremely warm years in the 18th century (Büntgen *et al.*, 2013 their fig. 3 and table S2). The May–June temperature in the years AD 1712, 1738, 1748, 1757, 1774 was higher than the mean temperature from 1961–1990. This may be recorded as the thermo-erosion of the ice in DIC.

Analyses of archival cave maps and photographs led to the conclusion that the ice level in DIC is recently in a steady-state condition (Clausen *et al.*, 2007). It is a cumulative effect of two processes: basal melting at the bottom of the ice block and simultaneous ice accumulation and partial melting on its top. The annual accumulation rate of ice on the top of the ice block has not been measured directly, although the accumulation rate of the ice was directly measured at some locations in DIC (Droppa, 1960; Halaš, 1989; Lalkovič, 1995; Tulis and Novotný, 2003). The results are as high as 15 cm per year. However, the measurements were conducted at specific sites, mostly those intensely fed with water. Tulis and Novotný (2003), on the basis of comparison of old cave maps with the present-day cave morphology, estimated the basal melting of cave ice in DIC as ca 1 cm/year. Accepting that the ice level is presently in steady-state (Vrana *et al.*, 2007), this value must reflect roughly the annual increment on the top of ice block. Thus, the recent accumulation rate is in the same range as the accumulation rate between ca 1210 and 1065 yr BP (ca AD 740–885). Basal melting of cave ice hardly depends on climatic conditions. Most probably, it results from the input of geothermal heat. The higher accumulation rate during the LIA must have resulted in the net accretion of the ice block. Thus, the empty space over the ice systematically decreased and the ice occupied more and more space in DIC.

Significance of fossil bats

Although fossil and subfossil bat remains were found in Slovakia, they were not dated radiometrically, with the exception of one bat, previously found in ice in DIC (sample KIA 23374 see Clausen *et al.*, 2007), but it is undetermined. Therefore, samples NIET-1 and NIET-2 represent the first determined and radiometrically dated bat remains form Slovakia. Among determined bats *Myotis mystacinus* (Kuhl), *M. brandti* (Eversmann) and *M. blythii* are known from Holocene deposits in Slovakia (Horáček, 1976; Ložek *et al.*, 1989). *Myotis alcathoe* (von Helversen et Heller) was distinguished relatively recently from *M. mystacinus* and *M. brandtii* (von Helversen *et al.*, 2001). Its remains have not been recognized in Holocene and older material, but some older findings of bats belonging to genera *Myotis* may represent this species.

All three species determined occur recently in Slovakia (Benda *et al.*, 2003; Uhrin *et al.*, 2010). *M. mystacinus/brandtii* and *M. blythii* are known to live presently in DIC (Uhrin, 1998; Bobáková, 2002a, b). Cold-adapted *M. mystacinus and M. brandtii* are dominant species there. Previously, bats that are known to be mainly cold-adapted, were recognized within the ice in DIC. They belong to *M. mystacinus*, *M. brandtii*, *Eptesicus nilssoni* (Keyserling et Blasius), *Plecotus auritus* (Linnaeus) (Obuch, 2012). However, the precise location of the samples in the ice section is not known.

The warmth-adapted *M. blythii* is noted mainly in the southern part of Slovakia, and DIC is near the northernmost margin of its distribution range (Uhrin *et al.*, 2008). It has been noted only once in the area located to the north of Slovakia (Piksa, 2006). The presence of the remains of this warmth-adapted bat in the fossil assemblage analyzed indicates that the ice that hosted this assemblage formed in a relatively warm climate, similar to the present-day climate. *M. blythii* avoids large and dense forests and semi-arid areas and so its remains indicate the presence of grassland (Arlettaz, 1996; Güttinger *et al.*, 1998). On the other hand, the bones of whiskered bats prove the existence of a more-or-less dense forest (von Helversen *et al.*, 2001). These notions are in line with the dating results that indicate the formation of the host ice during the MWP (see above).

Palaeoenvironmental implication of DIC ice

Generally, the accumulation rate of ice in DIC reflects late Holocene climate changes. The high accumulation rate during the LIA that is higher than in the present-day time and during the MWP. This resulted from suitable local conditions, which in turn were a consequence of global climate deterioration (colder and probably drier climatic conditions). Dating results of cave ice in other European caves indicate that in many of them there was ice formed during the

LIA. Spötl *et al.* (2014) noted the major ice accumulation during the LIA in Hundsalm Eishöhle und Tropfsteinhöhle (the Austrian Alps). Similarly, cave ice in Dachstein-Mammuthöhle (the Austrian Alps) originated during the LIA, which was proved by radiocarbon dating of a wood fragment, embedded in the lower part of the ice section (Mais and Pavuza, 2000).

The undisturbed accumulation of ice during the MWP is intriguing, since data from the Alps indicate 'starvation' of cave ice at that time, at least in some caves (e.g., Hundsalm Eishöhle und Tropfsteinhöhle – Spötl *et al.*, 2014). The above discrepancy can be explained by different climatic conditions in both regions, since the conditions during the MWP were geographically variable (Hughes and Diaz, 1994). Recent microclimatic studies in DIC indicate that an inflow of relatively warm water in the summer seasons can lead to substantial degradation of cave ice (Korzystka *et al.*, 2011; Strug, 2011). A dry summer can favour net accumulation, whereas a wet summer can lead to the destruction of cave ice. Thus, it can be hypothesized that the deficit of percolation water, caused by the relative dryness of summer seasons during the MWP, resulted in a net accumulation in DIC at that time. Unfortunately, the data concerning humidity during the MWP in the Carpathians are scarce and ambiguous. More precise data come from northern Romania, some 350–400 km SE of DIC. Geantă *et al.* (2012) noted a decline in beech pollen and an increase of oak and lime pollen in bat guano in Măgurici Cave (Someş Plateau). It can be a record of a drier climate during the MWP. This suggestion is in line with the pollen data from Fenyves-tető, a high-mountain mire in northern Romania, which indicates increasing dryness from ca 2240 cal. yr BP (Schnitchen *et al.*, 2006). Contradictory opinions were reached by Feurdean (2005), who studied peat deposits from the Gutaiului Mountains. Her data point to a change towards humid conditions from 2300 cal. yr BP. More complex data were obtained by Feurdean *et al.* (2015) from Tăul Muced mire. They indicated moderately dry summers between AD 800 and AD 900, succeeded by more humid conditions during the younger parts of the MWP. Thus, the existence of relatively dry conditions in summers during the MWP near DIC cannot be excluded. However, this hypothesis should be confirmed by independent methods. On the other hand, tree-ring width and cellulose $\delta^{13}C$ as well as testate amoeba and $\delta^{13}C$ point to wet conditions in the Alps between AD 900 and AD 1300 (Kress *et al.*, 2014 and van der Knapp *et al.*, 2011, respectively). This may explain the hiatuses in the ice sections in some Alpine caves.

Nonetheless, some local anomalies cannot be ruled out as a driving factor that resulted in the continuous accumulation rate in DIC. Similar anomalies can occur today. It is worth noting that DIC is an exceptional case worldwide, since it is one of a few ice caves, in which substantial ice recession is not observed.

CONCLUSIONS

1. Preserved frozen bats of the species *M. blythii* and the *M. mystacinus* morpho-group were found in the lowermost part of the ice section in DIC. They yielded radiocarbon dates 1266–1074 cal. yr BP (AD 684–876) and 1173–969 cal. yr BP (AD 777–981). The age of a previously found, undetermined bat is 1178–988 cal. yr BP (AD 772–962).

2. The dates obtained collectively indicate the accumulation of ice during the MWP. This is additionally confirmed by the presence of the relatively warmth-adapted *M. blythii*. The net accumulation of ice during this relatively warm period is hypothesized to have been possible owing to relative dry summer seasons at that time.

3. The ice started growing in DIC definitely earlier than the MWP since the phenomenon of basal melting was noticed in the cave. One can accept that the ice grew at least during the Dark Ages Cold Period which preceded the MWP, or plausibly even earlier.

4. The accumulation rate of ice during the LIA was higher than between ca 1625 yr BP and ca 1210 yr BP by a factor of 1.3 to 1.8, which is a record of climate deterioration.

Acknowledgments

Renata Jach is thanked for her help with the figures. The paper has greatly benefited from critical comments by Aurel Perşoiu and two other anonymous *ASGP* reviewers as well as from *ASGP* editors to whom the authors are greatly indebted.

REFERENCES

Arlettaz, R., 1996. Feeding behaviour and foraging strategy of free-living *Myotis myotis* and *Myotis blythii*. *Animal Behaviour*, 51: 1–11.

Behm, M., Dittes, V., Greilinger, R., Hartmann, H., Plan, L. & Sulzbacher, D., 2009. Decline of cave ice e a case study from the Austrian Alps (Europe) based on 416 years of observation. In: White, W. B. (ed.), *Proceedings of the 15th International Congress of Speleology, Kerrville, Texas, 19–26 July 2009, vol. 3*. International Union of Speleology, Kerville, pp. 1413–1416.

Bella, P., 2007. Morphology of ice surface in the Dobšiná Ice Cave. In: Zelínka, J. (ed.), *2nd International Workshop on Ice Caves. Proceedings*. Slovak Caves Administration, Liptovský Mikuláš, pp. 15–23.

Bella, P., 2008. The most important caves and abysses of the World Heritage. In: Jakál, J & Bella, P. (eds), *Caves of the World Heritage in Slovakia*. Slovak Caves Administration, Liptovský Mikuláš, pp. 139–154.

Benda, P., Ruedi, M. & Uhrin, M., 2003. First record of *Myotis alcathoe* (Chiroptera: Vespertillionidae) in Slovakia. *Folia Zoologica*, 52: 359–365.

Bobáková, L., 2002a. Quantitative and qualitative characteristics of bat fauna of the Dobšinská Ice Cave (E Slovakia). *Lynx*, 6: 47–58.

Bobáková, L., 2002b. Zimovanie netopierov v jaskynnom systéme Dobšinská ľadová jaskyňa – jaskyňa Duča. *Vespertilio*, 33: 245–248. [In Slovak.]

Bronk Ramsey, C., 2009. Bayesian analysis of radiocarbon dates. *Radiocarbon*, 51: 337–360.

Büntgen, U., Kyncl, T., Ginzler, C., Jacks, D. S., Esper, J., Tegel, W., Heussner, K. U. & Kyncl, J., 2013. Filling the Eastern European gap in millennium-long temperature reconstructions. *Proceedings of the National Academy of Sciences of the United States of America*, 110: 1773–1778.

Citterio, M., Turri, S., Bini, A. & Maggi, V., 2004. Observed trends in the chemical composition, $\delta^{18}O$ and crystal sizes vs. depth in the first ice core from the Lo Lc 1650 "Abisso sul Margine dell'Alto Bregai" ice cave (Lecco, Italy). *Theoretical and Applied Karstology*, 17: 45–50.

Clausen, H. B., Vrana, K., Bo Hansen, S., Larsen, I. B., Baker, J., Siggaard-Andersen, M. L., Sjolte, J. & Lundholm, S. C., 2007. Continental ice body in Dobšiná Ice Cave (Slovakia) – part II. – results of chemical and isotopic study. In: Zelínka, J. (ed.), *2nd International Workshop on Ice Caves. Proceedings*. Slovak Caves Administration, Liptovský Mikuláš, pp. 29–37.

Colucci, R. R., Fontana, D., Forte, E., Potleca, M. & Guglielmin, M., 2016. Response of ice caves to weather extremes in the southeastern Alps, Europe. *Geomorphology*, 261: 1–11.

Czernik, J. & Goslar, T., 2001. Preparation of graphite targets in the Gliwice Radiocarbon Laboratory for AMS ^{14}C dating. *Radiocarbon*, 43: 283–291.

Dietz, C., von Heleversen, O. & Nill, D., 2007. *Handbuch der Fledermäuse Europas und Nordwestafrikas*. Franckh-Kosmos Verlags, Stuttgart, 399 pp.

Droppa, A., 1960. *Dobšinská ľadová jaskyňa.* Vydavatelstvo Šport, Bratislava, 132 pp.

Fairchild, I. J. & Baker, A., 2012. *Speleothem Science. From Process to Past Environment*. Wiley-Blackwell, Chichester, 432 pp.

Feurdean, A., 2005. Holocene forest dynamics in northwestern Romania. *Holocene*, 15: 435–446.

Feurdean, A., Gałka, M., Kuske, E., Tantau, I., Lamentowicz, M., Florescu, G., Liakka, J., Hutchinson, S. M., Mulch, A. & Hickler, T., 2015. Last Millennium hydro-climate variability in Central–Eastern Europe (Northern Carpathians, Romania). *Holocene*, 25: 1179–1192.

Feurdean, A., Perşoiu, A., Pazdur, A. & Onac, B. P., 2011. Evaluating the palaeoecological potential of pollen recovered from ice in caves: A case study from Scărişoara Ice Cave, Romania. *Review of Palaeobotany and Palynology*, 165: 1–10.

Fuhrmann, K., 2007. Monitoring the disappearance of a perennial ice deposit in Merrill Cave. *Journal of Cave and Karst Studies*, 69: 256–265.

Gąsiorowski, M. & Sienkiewicz, E., 2010. The Little Ice Age recorded in sediments of a small dystrophic mountain lake in southern Poland. *Journal of Paleolimnology*, 43: 475–487,

Geantă, A., Tanţău, I., Tămaş, T. & Johnston, V. E., 2012. Palaeoenvironmental information from the palynology of an 800 year old bat guano deposit from Măgurici Cave, NW Transylvania (Romania). *Review of Palaeobotany and Palynology*, 174: 57–66.

Goslar, T., Czernik, J. & Goslar, E., 2004. Low-energy ^{14}C AMS in Poznań Radiocarbon Laboratory. *Nuclear Instruments and Methods* B, 223–224: 5–11.

Güttinger, R., Lustenberger, J., Beck, A. & Weber, U., 1998. Traditionally cultivated wetland meadows as foraging habitats of the grass-gleaning lesser mouse-eared bat (*Myotis blythii*). *Myotis*, 36: 41–49.

Haas, J. N., Richoz, I., Tinner, W. & Wick, L., 1998. Synchronous Holocene climatic oscillations recorded on the Swiss Plateau and at timberline in the Alps. *Holocene*, 8: 301–304.

Halaš, J., 1989. Le bilan thermique de la Grottte Glaciale de Dobšiná. *Slovenský Kras*, 27: 57–71. [In Slovak, with French summary.]

Hercman, H., Gąsiorowski, M., Gradziński, M. & Kicińska, D., 2010. The first dating of cave ice from the Tatra Mountains, Poland and its implication to palaeoclimate reconstructions. *Geochronometria*, 36: 31–38.

Hercman, H. & Pawlak, J., 2012. MOD-AGE: An age-depth model construction algorithm. *Quaternary Geochronology*, 12: 1–10.

Holzhauser, H., Magny, M. & Zumbühl, H. J., 2005. Glacier and lake-level variations in west-central Europe over the last 3500 years. *Holocene*, 15: 789–801.

Horáček, I., 1976. Review of Quaternary bats in Czechoslovakia. *Lynx*, 18: 35–58.

Hughes, M. K. & Diaz, H. F., 1994. Was there a "Medieval Warm Period", and if so, where and when? *Climatic Change*, 26: 109–142.

Jones, P. D., Osborn, T. J. & Briffa, K. R., 2001. The evolution of climate over the Last Millennium. *Science*, 292: 662–667.

Kern, Z., Fórizs, I., Kázmér, M., Nagy, B., Szántó, Z., Gál, A., Palcscu, L. & Molnár, M., 2004. Late Holocene environmental changes recorded at Ghetarul de la Focul Viu, Bihor Muntains, Romania. *Theoretical and Applied Karstology*, 17: 51–60.

Kern, Z., Fórizs, I., Pavuza, R., Molnár, M. & Nagy, B., 2011a. Isotope hydrological studies of the perennial ice deposit of Saarhalle, Mammuthöhle, Dachstein Mts, Austria. *Cryosphere*, 5: 291–298.

Kern, Z. & Perşoiu, A., 2013. Cave ice – the imminent loss of untapped mid–latitude cryospheric palaeoenvironmental archives. *Quaternary Science Reviews*, 67: 1–7.

Kern, Z., Széles, E., Horvatinčić, N., Fórizs, I., Bočić, N. & Nagy, B., 2011b. Glaciochemical investigations of the ice deposit of Vukušić Ice Cave, Velebit Mountain, Croatia. *Cryosphere*, 5: 485–494.

Kern, Z. & Thomas, S., 2014. Ice level changes from seasonal to decadal time-scales observed in lava tubes, Lava Beds National Monument, NE California, USA. *Geografia Fisica e Dinamica Quaternaria*, 37: 151–162

Korzystka, M., Piasecki, J., Sawiński, T. & Zelinka, J., 2011. Climatic system of the Dobšinská Ice Cave. In: Bella, P. & Gažík, P. (eds), *6th Congress International Show Caves Association Proceedings*. Slovak Caves Administration, Liptovský Mikuláš, pp. 85–97.

Kress, A., Hangartner, S., Bugmann, H., Büntgen, U., Frank, D. C., Leuenberger, M., Siegwolf, R. T. W. & Saurer, M., 2014. Swiss tree rings reveal warm and wet summers during medieval times. *Geophysical Research Letters*, 41: 1732–1737.

Lalkovič, M., 1995. On the problems of the ice filling in the Dobšiná Ice Cave. *Acta Carsologica*, 24: 313–322.

Laursen, L., 2010. Climate scientists shine light on cave ice. *Science*, 329: 746–747.

Ložek, V., Gaál, Ľ., Holec, P. & Horáček, I., 1989. Stratigraphy and Quaternary fauna of cave Peskő in the Rimava Basin. *Slovenský Kras*, 27: 29–56. [In Slovak, with English summary.]

Luetscher, M., Bolius, D., Schwikowski, M., Schotterer, U. & Smart, P. L., 2007. Comparison of techniques for dating of subsurface ice from Monlesi Ice Cave, Switzerland. *Journal of Glaciology*, 53: 374–384.

Luetscher, M., Jeannin, P. Y. & Haeberli, W., 2005. Ice caves as an indicator of winter climate evolution: a case study from the Jura Mountains. *Holocene*, 15: 982–993.

Luterbacher, J., Werner, J. P., Smerdon, J. E., Fernández-Donado, L., González-Rouco, F. J., Barriopedro, D., Ljungqvist, F. C., Büntgen, U., Zorita, E., Wagner, S., Esper, J., McCarroll, D., Toreti, A., Frank, D., Jungclaus, J. H., Barriendos, M., Bertolin, C., Bothe, O., Brázdil, R., Camuffo, D., Dobrovolný, P., Gagen, M., García-Bustamante, E., Ge, Q., Gómez-Navarro, J. J., Guiot, J., Hao, Z., Hegerl, G. C., Holmgren, K., Klimenko, V. V., Martín-Chivelet, J., Pfister, C., Roberts, N., Schindler, A., Schurer, A., Solomina, O., Von Gunten, L., Wahl, E., Wanner, H., Wetter, O., Xoplaki, E., Yuan, N., Zanchettin, D., Zhang, H. & Zerefos, C., 2016. European summer temperatures since Roman times. *Environmental Research Letters*, 11, 024001: 1–12.

Mais, K. & Pavuza, R., 2000. Hinweise zu Höhlenklima und Höhleneis in der Dachslein-Mammulhöhle (Oberösterreich). *Die Höhle*, 51: 121–125.

May, B., Spötl, C., Wagenbach, D., Dublyansky, Y. & Liebl, J., 2011. First investigations of an ice core from Eisriesenwelt cave (Austria). *Cryosphere*, 5: 81–93.

Obleitner, F. & Spötl, C., 2011. The mass and energy balance of ice within the Eisriesenwelt cave, Austria. *Cryosphere*, 5: 245–257.

Obuch, J., 2012. Assemblages of bats in deposits of the Dobšinská Ice Cave, Slovensky raj National Park, Slovakia. *Vespertilio*, 16: 205–210.

Perşoiu, A. & Pazdur, A., 2011. Ice genesis and its long-term mass balance and dynamics in Scărişoara Ice Cave, Romania. *Cryosphere*, 5: 45–53.

Pflitsch, A., Schörghofer, N., Smith, S. M. & Holmgren, D., 2016. Massive ice loss from the Mauna Loa Ice cave, Hawaii. *Arctic, Antarctic, and Alpine Research*, 48: 33–43.

Piksa, K., 2006. First record of *Myotis blythii* in Poland (Chiroptera: Vespertilionidae). *Lynx*, 37: 197–199.

Popa, I. & Kern, Z., 2009. Long-term summer temperature reconstruction inferred from tree-ring records from the Eastern Carpathians. *Climate Dynamics*, 32: 1107–1117.

Rachlewicz, G. & Szczuciński, W., 2004. Seasonal and decadal ice mass balance changes in the ice cave Jaskinia Lodowa w Ciemniaku, the Tatra Mountains, Poland. *Theoretical and Applied Karstology*, 17: 11–18.

Reimer, P. J., Bard, E., Bayliss, A., Beck, J. W., Blackwell, P. G., Bronk Ramsey, C., Buck, C. E., Cheng, H., Edwards, R. L., Friedrich, M., Grootes, P. M., Guilderson, T. P., Haflidason, H., Hajdas, I., Hatté, C., Heaton, T. J., Hoffmann, D. L., Hogg, A. G., Hughen, K. A., Kaiser, K. F., Kromer, B., Manning, S. W., Niu, M., Reimer, R. W., Richards, D. A., Scott, E. M., Southon, J. R., Staff, R. A., Turney, C. S. M. & van der Plicht, J., 2013. IntCal13 and Marine13 radiocarbon age calibration curves 0–50,000 years cal. BP. *Radiocarbon*, 55: 1869–1887.

Schnitchen, C., Charman, D. J., Magyari, E., Braun, M., Grigorszky, I., Tóthmérész, B., Molnár, M. & Szántó, Z., 2006. Reconstructing hydrological variability from testate amoebae analysis in Carpathian peatlands. *Journal of Paleolimnology*, 36: 1–17.

Schöner, W., Weyss, G. & Mursch-Radlgruber, E., 2011. Linkage of cave-ice changes to weather patterns inside and outside the cave Eisriesenwelt (Tennengebirge, Austria). *Cryosphere*, 5: 603–616.

Spötl, C., Reimer, P. J. & Luetscher, M., 2014. Long-term mass balance of perennial firn and ice in an Alpine cave (Austria): Constraints from radiocarbon-dated wood fragments. *Holocene*, 24: 165–175.

Stoffel, M., Luetscher, M., Bollschweiler, M. & Schlatter, F., 2009. Evidence of NAO control on subsurface ice accumulation in a 1200 yr old cave-ice sequence, St. Livres ice cave, Switzerland. *Quaternary Research*, 72: 16–26.

Strug, K., 2011. *Klimatyczne uwarunkowania rozwoju zjawisk lodowych w jaskiniach o odmiennych cechach środowiska*. Instytut Geografii i Rozwoju Regionalnego, Wrocław, 163 pp. [In Polish.]

Szukała, P., 2010. Zmiany bilansu masy lodu w Jaskini Lodowej w Ciemniaku (Tatry Zachodnie, Polska) z zastosowaniem technik trójwymiarowego modelowania jaskiń. In: Kotarba, A. (ed.), *Nauka a zarządzanie obszarem Tatr i ich otoczeniem, tom I.* Tatrzański Park Narodowy, Zakopane, pp. 145–149. [In Polish.]

Trofimova, E. V., 2006. Cave ice of Lake Baikal as an indicator of climatic changes. *Doklady Earth Sciences*, 410: 1087–1090.

Tulis, J. & Novotný, L., 1989. *System of Stratenská Cave*. Slovenská speleologická spoločnosť, Liptovský Mikuláš, 361 pp. [In Slovak, with English summary.]

Tulis, J. & Novotný, L., 2003. Zmeny zaľadnenia v Dobšinská ľadovej jaskyni. *Aragonit*, 8: 7–9. [In Slovak.]

Uhrin, M., 1998. Review of knowledge on bats (Mammalia: Chiroptera) of the Dobšinská Ice Cave - Stratenská Cave system. *Aragonit*, 3: 18. [In Slovak, with English summary.]

Uhrin, M., Benda, P., Obuch, J. & Danko, Š., 2008. Lesser mouse-eared bat (*Myotis blythii*) in Slovakia: distributional status with notes on its biology and ecology (Chiroptera: Vespertilionidae). *Lynx*, 39: 153–190.

Uhrin, M., Benda, P., Obuch, J. & Urban, P., 2010. Changes in abundance of hibernating bats in central Slovakia (1992–2009). *Biologia*, 65: 349–361.

van der Knapp, W. O., Lamentowicz, M., van Leeuwen, J. F. N., Hangertner, S., Leuenberger, M., Mauquoy, D., Goslar, T., Mitchell, E. A. D., Lamentowicz, Ł. & Kamenik, C., 2011. A multi-proxy, high-resolution record of peatland development and its drivers during the last millenium from the subalpine Swiss Alps. *Quaternary Science Reviews*, 30: 3467–3480.

von Helversen, O., Heller, K. G., Mayer, F., Nemeth, A., Volleth, M. & Gombkőtő, P., 2001. Cryptic mammalian species: a new species of whiskered bat (*Myotis alcathoe* n. sp.) in Europe. *Naturwissenschaften*, 88: 217–223.

Vrana, K., Baker, J., Clausen, H. B., Hansen, S. B., Zelinka, J., Rufli, H., Očkaik, Ľ. & Janočko, J., 2007. Continental ice body in Dobšiná Ice Cave (Slovakia) – Part I – Project and sampling phase of isotopic and chemical study. In: Zelínka, J. (ed.), *2nd International Workshop on Ice Caves. Proceedings*. Slovak Caves Administration, Liptovský Mikuláš, pp. 24–28.

Wilson, A. T., 1998. ^{14}C studies of natural ice. *Radiocarbon*, 40: 953–962.

Wołoszyn, B. W., 1970. The Holocene bat fauna (Chiroptera) from the caves of the Tatra Mountains. *Folia Quaternaria*, 35: 1–52. [In Polish, with English summary.]

Yang, S. & Shi, Y., 2015. Numerical simulation of formation and preservation of Ningwu ice cave, Shanxi, China. *Cryosphere*, 9: 1983–1993.

Yonge, C. I. & MacDonald, W., 1999. The potential of perennial cave ice in isotope palaeoclimatology. *Boreas*, 28: 357–362.

Zelinka, J., 2008. *Cave microclimate*. In: Jakál, J. & Bella, P. (eds), *Caves of the World Heritage in Slovakia*. Slovak Caves Administration, Liptovský Mikuláš, pp. 87–94.