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Karst processes from the beginning to the end: How can they be dated?

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Abstract

Determining the beginning and the end of the life of a karst system is a substantial problem. In contrast to most of living systems development of a karst system can be „frozen" and then rejuvenated several times (polycyclic and polygenetic nature). The principal problems may include precise definition of the beginning of karstification (e.g. inception in speleogenesis) and the manner of preservation of the products of karstification. Karst evolution is particularly dependent upon the time available for process evolution and on the geographical and geological conditions of the exposure of the rock. The longer the time, the higher the hydraulic gradient and the larger the amount of solvent water entering the karst system, the more evolved is the karst. In general, stratigraphic discontinuities, i.e. intervals of nondeposition (disconformities and unconformities), directly influence the intensity and extent of karstification. The higher the order of discontinuity under study, the greater will be the problems of dating processes and events. The order of unconformities influences the stratigraphy of the karst through the amount of time available for subaerial processes to operate. The end of karstification can also be viewed from various perspectives. The final end occurs at the moment when the host rock together with its karst phenomena is completely eroded/denuded. In such cases, nothing remains to be dated. Karst forms of individual evolution stages (cycles) can also be destroyed by erosion, denudation and abrasion without the necessity of the destruction of the whole sequence of karst rocks. Temporary and/or final interruption of the karstification process can be caused by the fossilisation of karst due to loss of its hydrological function. Such fossilisation can be caused by metamorphism, mineralisation, marine transgressions, burial by continental deposits or volcanic products, tectonic movements, climatic change etc. Known karst records for the 1st and 2nd orders of stratigraphic discontinuity cover only from 5 to 60 % of geological time. The shorter the time available for karstification, the greater is the likelihood that karst phenomena will be preserved in the stratigraphic record. While products of short-lived karstification on shallow carbonate platforms can be preserved by deposition during the immediately succeeding sea-level rise, products of more pronounced karstification can be destroyed by a number of different geomorphic processes. The longer the duration of subaerial exposure, the more complex are those geomorphic agents.

Owing to the fact that unmetamorphosed or only slightly metamorphosed karst rocks containing karst and caves have occurred since Archean, we can apply a wide range of geochronologic methods. Most established dating methods can be utilised for direct and/or indirect dating of karst and paleokarst. The karst/paleokarst fills are very varied in composition, including a wide range of clastic and chemogenic sediments, products of surface and subsurface volcanism (lava, volcaniclastic materials, tephra), and deep-seated processes (hydrothermal activity, etc). Stages of evolution can also be based on dating correlated sediments that do not fill karst voids directly. The application of individual dating methods depends on their time ranges: the older the subject of study, the more limited is the choice of method. Karst and cave fills are relatively special kinds of geologic materials. The karst environment favours both the preservation of paleontological remains and their destruction. On one hand, karst is well known for its richness of paleontological sites, on the other hand most cave fills are complete sterile, which is true especially for the inner-cave facies. Another problematic feature of karst records is the reactivation of processes, which can degrade a record by mixing karst fills of different ages.

Keywords: karst, speleogenesis, dating methods, geochronology

Principle: The time scale for the development of karst features cannot be longer than that of the rocks on which they form. (White 1988, p. 302)

1. Introduction

The beginning and the end of the life of living organisms (plants, animals) are really clear thresholds (insemination/pollination → death) that can be precisely determined and described. On the other hand, to establish the beginning and the end of the life of a karst system is a substantial problem. In contrast to most of living systems, the development of karst systems can be „frozen" (halted) and then rejuvenated, often for several times.

Fossilisation and rejuvenation of karst can be viewed according to thermodynamic principles

(Eraso 1989): when the external dissipation function of the system, which represents the velocity of entropy production, reaches a minimum, the system is in a stationary state – water circulation and its chemical potential for rock dissolution has ceased – and the karstification is interrupted. The introduction of new energy (hydraulic head) to the system may cause reactivation of karstification. Polycyclicity of karst formation is a typical feature (e.g., Panoš 1964; Ford and Williams 1989). The polygenetic nature of many karsts that evolved in several different steps should be stressed, too (Ford and Williams 1989), e.g., overprint of cold karst processes on earlier deep-seated/hydrothermal products, which themselves followed meteoric early diagenesis (e.g., Bosák 1997).

The dating of karst evolution poses philosophical problems, principally (1) the precise definition of the beginning of karstification, and (2) modes of preservation of any karstification products, recognising that karst rocks are more easily soluble than other rock types under specific conditions that differ with the individual lithologies (limestones, dolomites, gypsum, anhydrite, rock salt, quartzite). The role of preservation is very important because karstlands function as traps or preservers of the geologic and environmental past, especially of terrestrial (continental) history where correlative sediments are mostly missing, but also of evidences in the marine records (Horáček and Bosák 1989).

Karstification of the host rocks may start during their formation phases – diagenesis – changing the soft sediment to a consolidated rock shortly after deposition itself. Such karstification is a consequence of the emergence of part of a depocenter (sedimentary basin) and the introduction of meteoric water to the diagenetic system. The formation of a freshwater lens and a halocline zone related to the surface relief and sea-level changes is the result. The early stages of origin of dissolutional (karst) porosity by meteoric diagenesis in carbonate rocks have been described in numerous sedimentological and paleokarst studies (e.g., Longman 1980; James and Choquette 1984; Tucker and Wright 1990; James and Choquette, Eds. 1988; Wright, Esteban and Smart, Eds. 1991; Moore 1989, 2001). Some authors suppose karst to be merely the facies of meteoric diagenesis (Esteban and Klappa 1983).

The evolution of a karst depends especially on the time available for processes to operate and on the geographical and geological conditions of rock exposures. The longer the time available, the higher the hydraulic gradient and the larger the quantity of solvent water entering the system, the more evolved will be the karst in all its modes of occurrences (exo- and endokarst). In general, we can state that the kind of stratigraphic discontinuities, i.e. intervals of nondeposition (disconformities and unconformities; see Esteban 1991), directly influences the intensity and extent of karstification. The higher the order of discontinuity under study, the bigger are the problems to be expected when dating the processes and events.

The end of karstification can also be viewed from various perspectives. An undisputed end of karstification occurs at the moment when host rock together with its karst phenomena is completely eroded/denuded, i.e. at the end of the karst cycle *sensu* Grund (1914; see also Cvijić 1918). In such a case, nothing is left to be dated. Karst forms of individual stages of evolution (cycles) can also be destroyed by other, non-karst processes of erosion or by the complete filling of epikarst and burial of karst surfaces by impermeable sediments, without the necessity of destroying an entire sequence of karst rocks (the cycle of erosion of Davis 1899; see also Sawicki 1908, 1909). Temporary and/or final interruption of karstification can be caused by fossilisation due to the loss of the hydrological function of the karst (Bosák 1989, p. 583). The fossilisation can be caused by metamorphism, mineralisation, marine transgressions, burial by continental deposits or volcanic products, tectonic movements, climatic change etc. (see Bosák 1989).

The principal question in this paper is: *Can we date karst processes at all?* The answer is given at the end. The paper deals mostly with karst in carbonate rocks, although the geochronologic methods and some of approaches reviewed are universal.

Unconformities: the time frame

The beginning and the end of karst is clearly associated with conformities, unconformities and disconformities. Esteban (1991) in an excellent review following the sequence stratigraphic approach outlined the role of nondepositional events (stratigraphic discontinuities) in karst evolution. Different ranks of stratigraphic discontinuity represent the differing time gaps in deposition that have been available for dissolution (karstification; see also Moore 1991, pp. 247-264).

The stratigraphic discontinuity (gap, lacuna) represents the chronostratigraphic interval(s) missing through nondeposition (hiatus) and/or lithostratigraphic interval(s) missing through erosional truncation. Excluding conformities, Esteban (1991) proposed classification of unconformities into (1) single (SUK) and (2) composite (CUK), both with measurable stratigraphic gaps. Conformities have no measurable stratigraphic gap and correspond to bedding planes or parasequence boundaries. The single unconformity represents a stratigraphic gap equivalent to a sequence boundary and the composite one is formed by the stacking or superposition of single unconformities (Esteban 1991, p. 92). A hierarchy of stratigraphic discontinuities was proposed, too (Fig. 1). Most (paleo)karsts are composite unconformities, representing long timespans without deposition.

Stratigraphy of karst

The order of unconformities influences the stratigraphy of the karst due to the time involved in subaerial processes (Table 1). There are two general systems of the karst stratigraphy based on: (1) the carbonate sedimentological/sequence stratigraphic approach (Choquette and James 1988), and (2) general karst models (Bosák, Ford and Głazek 1989).

Choquette and James (1988) distinguished: (1) *depositional karst*, (2) *local karst*, and (3) *interregional karst*. They noted, that to distinguish the products of local and interregional karsts may be difficult in some cases. Esteban (1991) stressed that the depositional karst of Choquette and James (1988), which is associated with parasequence boundaries (single unconformities) reflects a *Caribbean model* of karst development, while interregional karst resulting from complex evolution producing composite unconformities represents the *general* (non-Caribbean) *model* of karst.

The *Caribbean model* (Esteban 1991, p. 93) is characterised by brief exposure time, unstable carbonate mineralogy, shallow burial, minor tectonics, minor deep (freshwater) phreatic zone, with primary and fabric-selective porosities predominant, restriction to tropical to semi-arid environments, diffuse recharge-diffuse flow only, affected by mixing marine zone processes but not by hydrothermal mixing. However, geothermal-driven convection of groundwater has been detected in some Caribbean-type of settings (e.g., Rougerie and Wauthy 1993).

The *General model* (Esteban 1991, p. 93) is characterised by longer exposure time, stable mineralogy, deep burial, one or several tectonic events, an important deep phreatic zone, secondary and fracture porosities predominant, a wider range of climatic environments, confluent recharge, pipe and confined flow, absence of mixing marine zone effects and presence of hydrothermal mixing.

Local karst forms when part of a carbonate shelf is exposed, usually because of tectonism, drops in sea level or synsedimentary block tilting. Depending on the length of time involved, the effects of exposure can vary from minor to extensive with the development of exo- and endokarst (Choquette and James 1988, pp. 16-17).

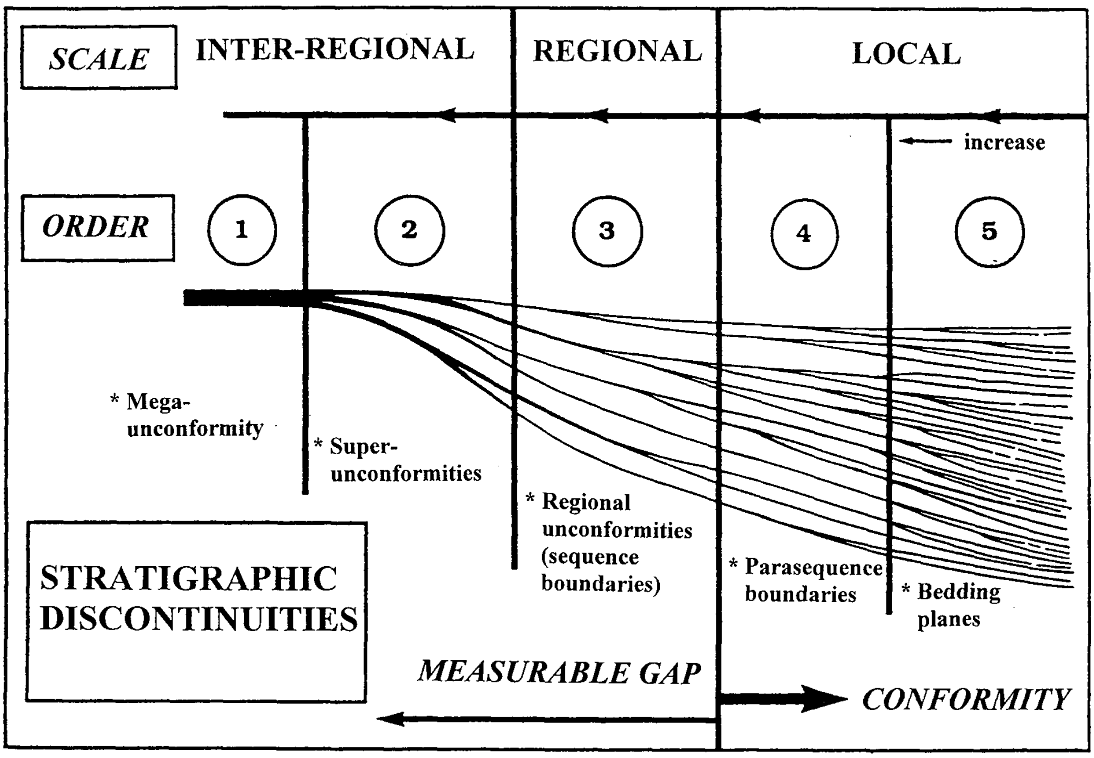

Fig. 1. Hierarchy of stratigraphic discontinuities (modified after Esteban 1991).

TABLE 1

Stratigraphic discontinuities, time gaps (modified after Esteban 1991) and stratigraphy of karst

<table>
<tr><th colspan="2" rowspan="2">STRATIGRAPHIC DISCONTINUITIES</th><th rowspan="2">ORDER</th><th colspan="2">TIME GAP SCALE</th><th rowspan="2">CORRESPONDING STRATIGRAPHIC UNITS</th><th colspan="2">STRATIGRAPHY OF KARST</th></tr>
<tr><th>Ma</th><th>Chrono-stratigraphy</th><th>James & Choquette, Eds. 1988</th><th>Bosák et al., Eds. 1989</th></tr>
<tr><td>UNCONFORMITIES
COMPOSITE</td><td>E. UNCONFORMITY
MEGAUNCONFORMITY
SUPERUNCONFORMITY SET
SUPERUNCONFORMITY</td><td>1

2</td><td>200
>60
30
4-12</td><td>ERATHEM
SYSTEM
SERIES
STAGE</td><td>MEGASEQUENCE
SUPERSEQUENCE SET
SUPERSEQUENCE</td><td>INTER-REGIONAL KARST</td><td>KARST PERIOD</td></tr>
<tr><td>SINGLE</td><td>REGIONAL UNCONFORMITIES (sequence boundaries)</td><td>3</td><td>~1</td><td>BIOZONE</td><td>DEPOSITIONAL SEQUENCE</td><td rowspan="2">LOCAL KARST</td><td rowspan="4">KARST PHASE</td></tr>
<tr><td rowspan="3">CONFORMITIES</td><td>SYNTECTONIC UNCONFORMITIES</td><td>3-4</td><td>$0.0X^{-1}$</td><td>Variable</td><td></td></tr>
<tr><td>BOUNDARY OF SHOLAING CYCLES</td><td>4</td><td>0.0X</td><td rowspan="2">Not recognisable</td><td>PARASEQUENCE</td><td rowspan="2">DEPOSITIONAL KARST</td></tr>
<tr><td>BEDDING PLANE</td><td>5</td><td>0.00X</td><td>BED</td></tr>
</table>

Interregional karst is much more widespread, is related to major eustatic-tectonic events, and results in karst terranes that may exhibit profound erosion, a wide variety of karst features, and deep, pervasive dissolution (Choquette and James 1988, p. 17).

Depositional karst forms as a natural consequence of sediment accretion at and around sea level. It is to be expected within the sediment packages that typify carbonate platforms. It is most commonly associated with meter-scale depositional cycles (Choquette and James 1988, p. 16).

Bosák, Ford and Głazek (1989) distinguished between: (1) the *karst phase*, and (2) the *karst period*. The connection to individual types of unconformities clearly proves the temporal relationships between all types of the karst, which may be mutually correlated (Table 1).

Karst period defines long-lasting times of groundwater circulation and continental weathering, which were terminated by an ensuing marine transgression. They are recognised by higher order unconformities or disconformities (= *interregional karst* of Choquette and James (1988). Their karst features can usually be divided into several generations (→ karst phases). Głazek (1989a) defined the tectonic conditions for karst periods as being induced by orogenies. Those lengthy periods are caused by the post-collisional uplift of orogens and their fringes. The periods are marked by unconformities and disconformities over broad areas and need not to be confined to individual modern continents. These long periods display diachronicity and many lesser phases. They are longest in duration and most complex at former mountain crests and become gradually shorter on the former mountain slopes and their broad fringes along adjacent continents. These periods result from major changes of plate motion patterns and they divided structural complexes corresponding to orogenic cycles (Głazek 1973).

A karst *phase* is caused by a geodynamic or major climatic change, e.g., uplift or downwarping, sea-level change, a phase of permafreezing, etc. (Bosák, Ford and Głazek 1989). From the tectonic point of view, Głazek (1989a) distinguished two kinds of karst phases: (1) represented as unconformities within the limited areas of one past shallow marine platform and its continental fringes, or of one continent created by the collision of two plates (= *local karst* of Choquette and James 1988); and (2) disconformable or paraconformable surfaces resulting from glacial-eustatic fluctuations of sea level or from local tectonic events (= *depositional karst* of Choquette and James 1988).

Interregional (paleo)karst and products of karst periods can be linked with composite unconformities of the 1st and 2nd orders *sensu* Esteban (1991). Such products can be correlated over extensive regions, e.g., post-Kaskadia and post-Variscan karstifications in North America and Europe, respectively (Głazek 1989a). Local (paleo)karst and products of type 1 of karst phases (*sensu* Głazek 1989a) are common products during single unconformities and syntectonic unconformities, i.e. of the 3rd order.

Karst forms created during the 4th and 5th order unconformities (conformities) correspond to depositional (paleo)karst and to Type 2 karst phases.

Karst record

The principal differences between the Caribbean karst model and the general karst model are concerned with exposure time. The former is associated with brief exposures to subaerial agents, i.e. with stratigraphic discontinuities of 3rd to 5th order with durations of 0.00X to about 1 Ma, the latter with lengthy exposures corresponding to stratigraphic discontinuities of 2nd and 1st order; i.e. with times of $X0^0$ to $X0^2$ Ma (Table 1).

The karst record of 1st and 2nd order stratigraphic discontinuities on the Eastern European Platform and epi-Variscan Central European Platform in Poland was identified by Głazek, Dąbrowski and Gradziński (1972) and Głazek (1973, 1989a). It encompasses a maximum of 50 to 60 % of the geological time elapsed since deposition of the rocks (Fig. 2). Analysis of the Bohemian Massif (epi-Variscan Platform; Bosák 1987, 1997; Table. 2, Fig. 3) showed that 12 to 45 % of geological time since the regression of Paleozoic seas in the Upper Devonian/Lower Carboniferous is in such records and 55 to 88 % of time is not recorded in the preserved marine or continental sequences (Bosák 1987).

These two examples of platform areas differ in the time recorded in the subsequent cover sediments. The Bohemian Massif is a relatively young body resulted from amalgamation of individual terranes during the Variscan Orogeny. Since that time uplift has prevailed over subsidence as a consequence of the tectonic stress caused by the Alpine Orogeny in its foreland. Platform sediments are rather rare there (Upper Jurassic and Upper Cretaceous regional transgressions, several minor Oligocene and Miocene transgressions covering only margins of the massif; see Fig. 3). The Polish territory is composed of slightly older elements in a different geotectonic setting, and the geologic structure is little affected by younger orogenies. Platform cover is developed more continuously and individual stratigraphic discontinuities are shorter. Therefore, there is a significant difference in the preserved record of time in the two regions, i.e. 12-45 % vs. 50-60 %. Some old cratonic units can be nearly completely without any platform cover (e.g., Scandinavian Shield), partly as a consequence of glacial isostasy. In such terranes, the time recorded can represent less than 10 %. On some recent and fossil carbonate platforms, time recorded in sediments represents only 5 to less than 10 % (Great Bahama Bank, Devonian carbonate platform on Moravia; Bosák et al. in print).

It can be readily asserted that the shorter the time available for karstification, the greater is the probability of preservation of the karst phenomena in the stratigraphic record. While products of short-lived karstification on shallow carbonate platforms can be preserved by deposition during the sea-level rise following immediately after, products of more pronounced karstification may be destroyed by a variety of geomorphic processes. The longer is the duration of subaerial exposure, the more complex are those geomorphic agents. Further, individual long periods of subaerial exposure (stratigraphic discontinuities of the 1st and 2nd orders – karst periods) may coalesce, being separated only by a short interruption (e.g., marine transgression/ ingression).

TABLE 2

Review of temporal data for the evolution of the Bohemian Massif since the Paleozoic regression (after Bosák 1987, 1997)

Regional geological unit	Duration since regression (Ma)	Record preserved (Ma)	Record in continental deposits (Ma)	Record (%)	Gap without record (%)
Moldanubicum	375	45	45	12	88
Bohemicum	375	48	36	13	87
Saxothuringicum	420	52	40	12	88
Brunovistulicum					
a. in outcrops	320	75	36	23	77
b. covered by Carpathian Foredeep	320	100-145	2	31-45	69-55

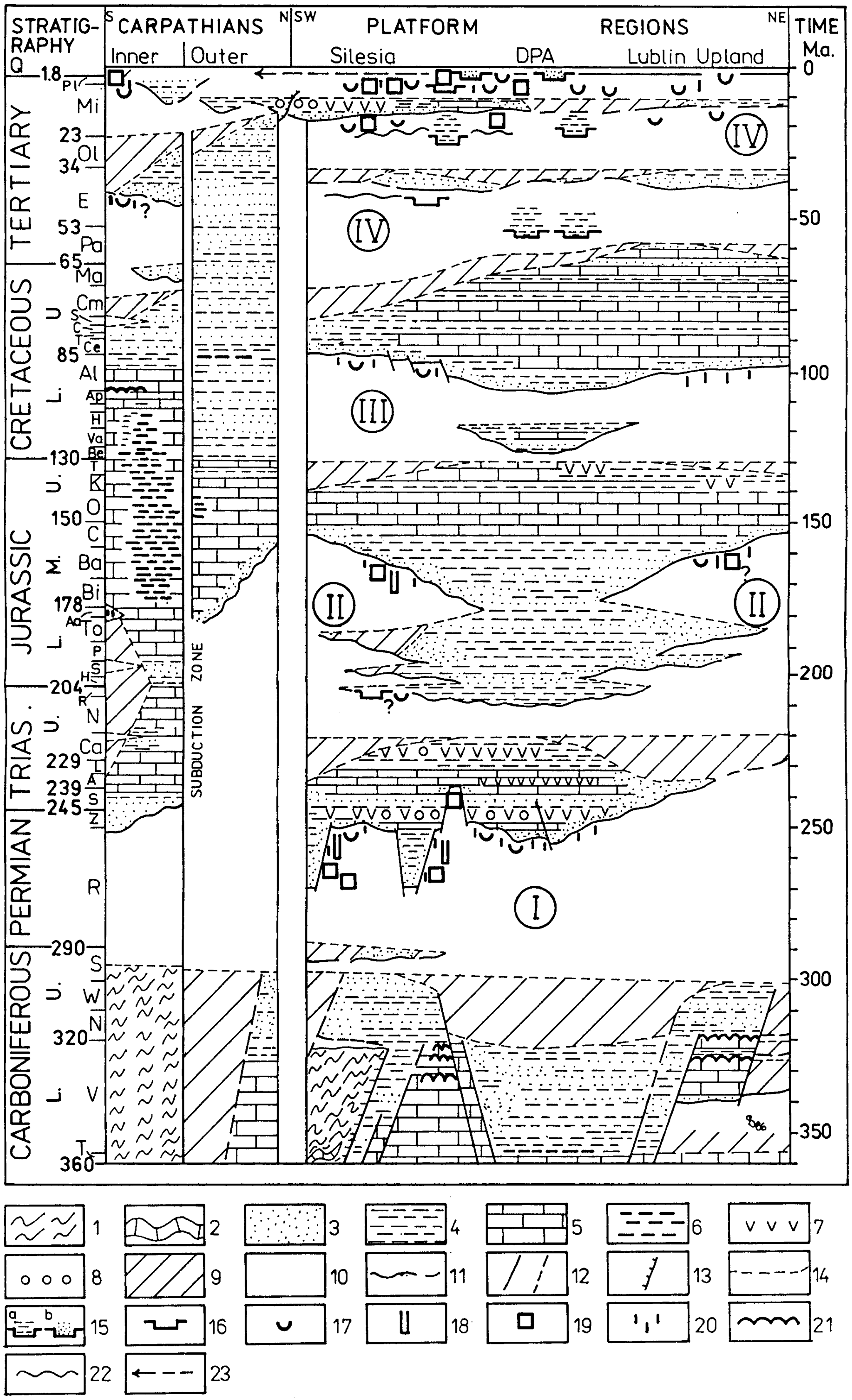

Fig. 2. Time distribution of paleokarst phenomena and sediments in Poland (from Głazek 1989b; with permission). Metamorphosed basement: 1 – silicate rocks, 2 – marble lenses; Sedimentary rocks: 3- psammites and psephites, 4 – silts, clays, marls, 5 – carbonates, 6 – deep-sea carbonate-silicate, 7 – sulphates, 8 – salts, 9 – unknown deposits (eroded), 10 – subaerial degradation; Boundaries: 11 – unconformable cover, 12 – synsedimentary faults, 13 – synsedimentary overthrusts, 14 – supposed limits of deposition, 15 – subrosion depressions with fills (a. brown coal, b. drift deposits), 16 – poljes, 17 – sinkholes, 18 – shafts, 19 – caves, 20 – minor solution forms, 21 – syngenetic caves, 22 – karst corrosion surfaces, 23 – maximal extent of Pleistocene glaciers, I to IV – periods of karstification.

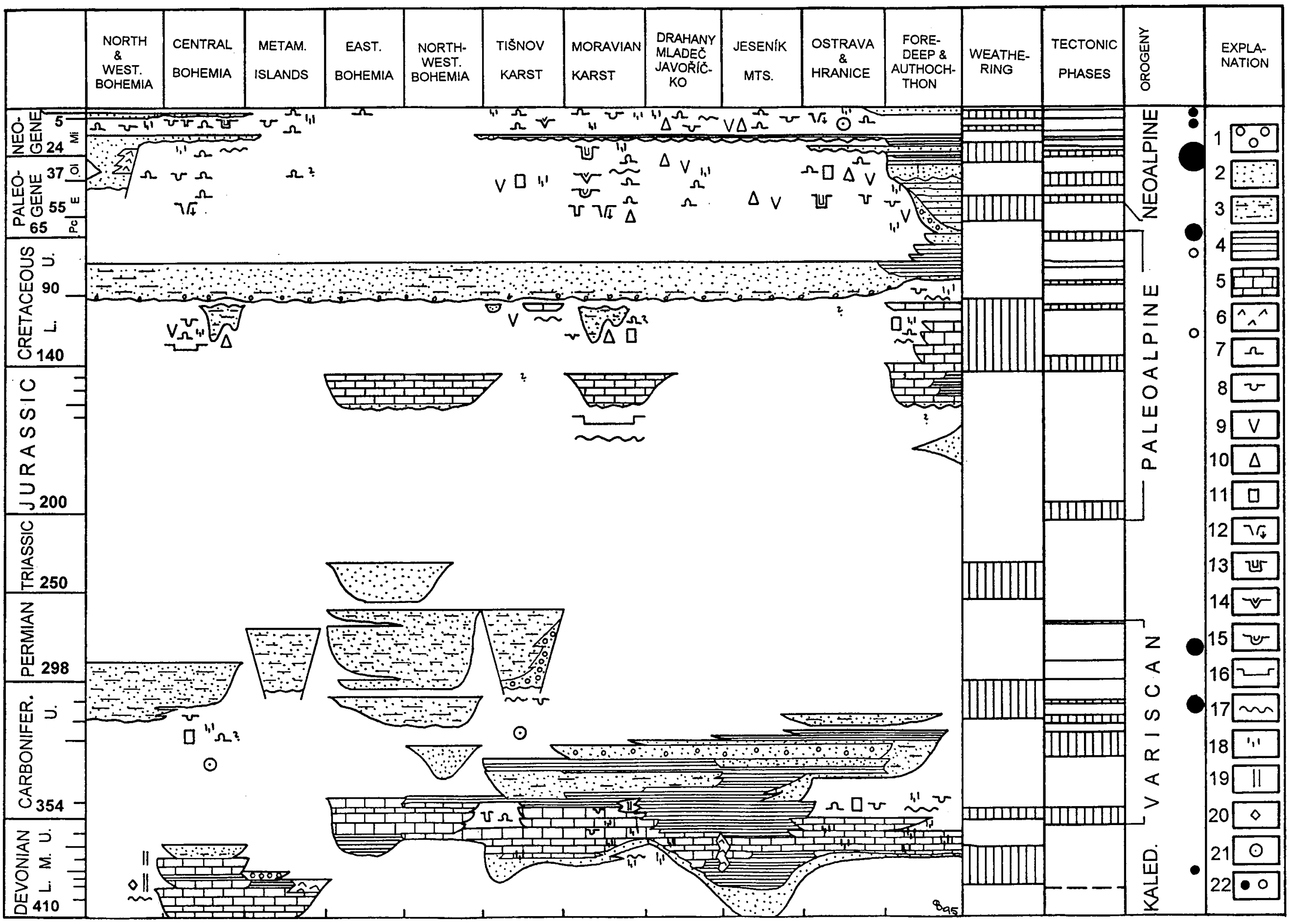

Fig. 3. Distribution of paleokarst and sediments in selected sections of the Bohemian Massif (simplified and schematized; modified after Bosák 1997). Lithology: 1 – conglomerates, 2 – sandstones, 3 – lithologically variable siliciclastics (redbeds, alternation of sandstones, siltstones, sandstone, etc.), 4 – shales, 5 – carbonate rocks, 6 – volcanics and volcanoclastic rocks; Karst forms: 7 – caves, 8 – dolines, 9 – geological organs, 10 – karst cones, 11 – karst inselbergs, 12 – collapse shafts, 13 – canyons, 14 – V-shaped valleys, 15 – U-shaped valleys, 16 – poljes and large karst depressions, 17 – corrosional surfaces, 18 – karren and minor solution forms, 19 – neptunian dykes, 20 – meteoric diagenetical vugs, 21 – hydrothermal karst, 22 – volcanic activity, black - Bohemian Massif, circle - Outer Western Carpathians adjacent to the Bohemian Massif, circle diametre approximately covers the time-span of volcanic activity.

Products of paleokarst evolution are best preserved directly beneath a cover of marine or continental sediments, i.e. under the deposits, which terminate the periods or phases of karstification. The longer the duration of the stratigraphic gap the more problematic is the precise dating of the paleokarst, unless it can be chronostratigraphically proven. Therefore, the ages of particular paleokarsts have been assigned mostly to times shortly before the termination of the stratigraphic gap (Bosák 1997). This fact can be easily illustrated in the Bohemian Massif for pre-Cenomanian age paleokarst, for pre-Callovian in the Moravian Karst or for Westphalian/Stephanian in central Bohemia (see Fig. 3). An identical situation occurs in Poland (Głazek 1989b; see Fig. 2)

Some processes can destroy karst features in relatively short time, leaving planated surfaces with little or no traces of previous karstification, e.g., the effect of marine transgressions. This can be illustrated from recent karst in the coastal zone of Palawan Island (Philippines) and the Lower Devonian of the Koněprusy area, Czech Republic. On Palawan, Longman and Brownlee (1980) described wave and surf action destroying or undercutting recent shore cliffs up to 30 m high that were composed of highly karstified limestones with dense networks of pinnacle karren, leaving only a flat abrasion platform with only rare relics of truncated solution fissures and sinkholes in their place. An identical situation is detected at the boundary between Koněprusy Limestones (Pragian) and Suchomasty Limestones (Dalejan, Lower Devonian) at Koněprusy. The truncation plane, which is nicely exposed in Koněprusy Caves, is smoothed by marine abrasion and shows no trace of karst, although the limestones contain distinct traces of meteoric diagenesis and the formation of neptunian dykes correlated with the hiatus, which lasted about 5-6 Ma.

Minimum time for speleogenesis

The evolution of a conduit is rather complicated set of events facing numerous critical thresholds (for summary see White 1988 and Ford & Williams 1989). At the present time, two phases of speleogenesis are generally accepted: (1) initiation – initial enlargement of a fracture to a critical size, and (2) enlargement – growth of a protoconduit to full conduit size (White 1988, p. 287). The initial fracture permeability and/or rock porosity has connected apertures on the order of 50 -500 μm and the diameter of a dissolutional proto-conduit reaches 5-15 mm (White 1988; Ford and Williams 1989). At diameters of 0.5 to 5 cm there is a kinetic breakthrough (Dreybrodt and Gabrovšek 2000) and flow may change from laminar to turbulent (White 1988, p. 291; Ford and Williams 1989).

The duration of a typical *initiation phase* was calculated to be about 3-5 ka (White 1988) based on experiments of Howard and Howard (1967) and calculations of Palmer (1981). They stated that the maximum dissolution rate is 0.14 $m.a^{-1}$. Palmer (1991) calculated the initiation phase to minimum of 10 ka under favourable conditions. Dreybrodt and Gabrovšek (2000) estimated the duration of the initiation (gestation) phase for realistic cases from 1 ka to 10 Ma. The time depends critically on the length and the initial width of the fracture.

The *enlargement phase*, i.e. the time in which protoconduit enlarges into full size (of 1-10 m or more) is expected to be 5 - 20 ka up to 100 ka in many geologic settings (White 1988). Ford and Williams (1989, p. 166) suggested that conduits can expand to diameters of 1-10 m in a few thousands of years (see also Palmer 1991), or even in a few hundreds years in high relief, wet terrains. Palmer (1991) calculated the maximum wall retreat to 0.01-0.1 cm/a in a typical meteoric groundwater cave. For hydrothermal caves, times on the order of 10^5 to 10^6 years are required to produce caves of traversable size (Palmer 1991, p 18). Data of Ford (1980) and Palmer (1984) suggest that an extension time of 10 to 100 ka per kilometre of the conduit may have prevailed in a majority of karst settings. White (1984) obtained an extension rate of 3-5 ka per kilometre. Dreybrodt and Gabrovšek (2000) estimated the velocity of enlargement of a conduit under phreatic conditions to about 200 mm/ka, so a phreatic passage of 30 m diameter can be developed within 100 ka. Of course, all those estimates are only illustrative as the velocity of speleogenesis is affected by numerous thresholds (see e.g., White 1988) and agents including geologic conditions (lithology, primary and secondary porosity), climatic conditions (temperature, precipitation, water volumes), hydrochemical conditions (concentration and kind of solvent agents), etc.

Theoretical assumptions have been proven by field observations. Mylroie and Carew (1986, 1987) dated the origin of Lighthouse Cave (San Salvador Island, Bahamas) between 85 ka (cementation of eolianite host rock) and 49 ka (U-series datum from a stalagmite), i.e. 36 ka available for the cave formation along the halocline. Numerous data from North America or Ireland indicate the post-glacial origin of caves perfectly adjusted to recently deranged surface landscapes and hydrologic regimes, i.e. caves developed during the last 8-15 ka (e.g., Mylroie and Carew 1986, 1987; White 1988; Ford and Williams 1989).

Determining the age of a cave is a problem because the dating is based on cave deposits (both clastic and chemogenic). In most cases we are able to date only the last few events of cave filling. Cases where the original syngenetic cave fill is preserved are rare, e.g., phreatic clays and silts, hydrothermal speleothems quasi-synchronous with phreatic speleogenesis. The dynamic character of karst results in repeating infilling and excavation of cave fills, under differing specific conditions. For example, in the Czech Karst only young Middle and Late Pleistocene deposits are preserved in the caves, with older Quaternary and pre-Quaternary fills found in some vertical corroded fissures as result of sequences of cave fills and exhumations (Ložek and Skřivánek 1965). In the Moravian Karst (Czech Republic), the situation is very similar (Kadlec et al. 2001), although the principal caves in both karst regions are at least of Early Miocene age. Complex watertable caves with pronounced flood histories offer only the age of the last cave fill episode. In Slovenia, Trhlovca and Divaška Caves (Classical Karst) contain sedimentary fill about 0.7 to 1.1 Ma old (Brunhes/Matuyama boundary and Jaramillo subchron; Bosák et al. 2000; Pruner and Bosák 2001), representing the last flood-derived fills. The system of Domica-Baradla Cave (Slovakia-Hungary), although pre-Pliocene in age, is filled only by the Late Pleistocene sediments (magneto-stratigraphy and U-series dating; Pruner and Bosák 2001 and yet unpublished data of Bosák/Pruner and D.C.Ford teams). So the age of the cave itself (void within the rock) is very far from obtained dates.

Inception: The start of a cave?

The preceding discussion has summarised the characteristic time scale for the development of a conduit. The same scale (i.e. 10 to 100 ka) is demanded for the development of a surface landform (White 1988, p. 304). Nevertheless, for caves concepts of legacy karst (V.P.Wright 1991; Wright and Smart 1994) or inception (Lowe 1999) have also been proposed that suggest that there exist pre-requisities guiding at least some speleogenesis.

Legacy karst according to V.P. Wright (1991) and Wright and Smart (1994) refers to dissolution occurring at the present or in the past whose distribution is controlled by an earlier (paleo)karst system. *Inception* according to Lowe (1999, 2000) is limited to a minor subset of all stratigraphic partings, those which dominate initially, imprinting incipient guidance for the later cave development. The weaknesses are imprinted within carbonate sequences during or soon after diagenesis, and certainly pre-tectonically. According to Waltham (2000) the inception horizon is a feature within the limestone structure that is a favourable site for the critical first phase of cave enlargement. The feature may be physical or chemical – a fracture, a mineralised fault, a shale bed containing pyrite, or a contrast in limestone chemistry. It is the initial inception stage and not the subsequent development stages, that provides the key to understanding where caves lie. The inception is a part of the initiation phase of cave formation (Lowe and Gunn 1997). It can be commented, that long ago, Ford (1971) stated that some planes or contacts are preferred *locii* of initiation in some caves. In that view the concept of inception, which states the same, seems to be rather complicated.

Taking these concepts into account, reflecting the polycyclic and polygenetic nature of much karst, we are facing a serious problem: how to define the age of origin of caves (protoconduits)? We have two possibilities of approach: (1) to accept all previous paleokarst features as the beginning of speleogenesis (even meteoric diagenesis), or (2) to accept only the result of the last speleogenetic phase (where it is the phase that created the known cave), ignoring all previous events. The second option seems to offer some problems in specific settings.

For example, the origin of the Lighthouse Cave (San Salvador Island, Bahamas; Mylroie and Carew 1986, 1987) was as a single event piece of speleogenesis in upper Pleistocene rocks (~125,000 years in age) without any legacy karst; there is no problem to place the beginning of speleogenesis after that age. On the other hand, speleogenesis in the Koněprusy region (Czech Republic; Bosák 1996, 1997, 1998) identified by the analysis of hundreds of cored boreholes in Lower Devonian limestones indicate that each succeeding phase of karstification utilised previously karstified („prepared“) space, starting with Lower/Middle Devonian diagenetic (mostly meteoric) vuggy porosity and neptunian dykes, followed by late Variscan hydrothermal karst (Carboniferous/Permian), Lower/Middle Cretaceous karstification and finally by a complex set of confined hydrothermal/cold karstification during Paleogene/Miocene time = the complex and prolonged history of polycyclic and polygenetic karst with many interruptions in formation and many changes of geologic and climatic conditions. Where is the beginning of speleogenesis to be dated?

Geochronologic methods

Owing to the fact that unmetamorphosed or only slightly metamorphosed karst rocks have existed since the Proterozoic, we are facing the wide range of application of geochronologic methods. The oldest karst forms with caves and cave deposits are known from Early Proterozoic of Transvaal, South Africa (2.2 Ga; Martini 1981). Karst breccias of Archean age are known in the Canadian Shield (D.C. Ford, pers. comm. 2002). Somewhat younger are paleokarst surfaces in Canada (Belcher Island – 1.7 Ga, Ontario and Quebec – 1.4 Ga; Ford 1989). Upper Proterozic karst is also known from several locations on old cratons and platforms, e.g., in China (Zhang Shouyue 1989), Russia (Tsykin 1989) or Australia (Rowlands et al. 1980).

Most of the methods outlined below can be utilised for direct and/or indirect dating of karst and paleokarst processes. Karst/paleokarst fills are highly variable in origin and composition, including a wide range of clastic and chemogenic sediments, products of surface and subsurface volcanism (lava, volcaniclastic materials, tephra), and deep-seated processes (hydrothermal activity, etc). During burial, paleokarst forms can be cut or penetrated by products of younger deep-seated processes (volcanic or hydrothermal – ore – veins). Evolutionary karst stages can be based also on dating of correlative sediments, which do not fill karst voids directly, i.e. glacial deposits, river terraces, eolian and lacustrine sediments, marine deposits and fossils. Certain dating methods cannot be used for karst events at all, especially those requiring magmatic and/or metamorphic lithologies as suitable materials.

Colman and Pierce (2000) reviewed the range of geochronologic methods for the Quaternary period. Their conclusions can be adapted also for older chronologic units. The methods are grouped into six categories: (1) *sidereal* (calendar or annual) methods, which determine calendar dates or count annual events; (2) *isotopic* methods, which measure changes in isotopic composition due to radioactive decay and/or growth; (3) *radiogenic* methods, which measure cumulative effects of radioactive decay, such as crystal damage and electron energy traps; (4) *chemical and biological* methods, which measure the results of time-dependent chemical or biological processes; (5) *geomorphic* methods, which measure the cumulative results of complex, interrelated, physical, chemical, and biologic processes on the landscape; and (6) *correlation* methods, which establish age equivalence using time-independent properties. Results of dating can be classified into four groups as follows: *numerical-age*, *calibrated-age*, *relative-age*, and *correlated-age* (Colman and Pierce 2000, p. 3). They also proposed to abandon the term *absolute date* in favour of *numerical date*.

The application of individual dating methods depends on their timespans. In general, we can state that the older is the subject of our study, the more limited are the methods of dating available. The nature of geologic materials to be dated represents another threshold. Not all geologic materials are suitable for numerical dating. On the other hand, most of materials are suitable to attempt correlated-age.

Karst and cave fills are relatively special kinds of geologic materials. The karst environment favours both the preservation of paleontological remains and their destruction. On one hand, karst is well known for its wealth of paleontological sites (see e.g., Horáček and Kordos 1989), on the other hand most cave fills are completely sterile, especially for the inner-cave facies. Another problematic feature of karst records is that there may be reactivation of processes, which degrades the record into an unreadable form, often mixing karst fill of different ages (collapses, redepositions, etc., e.g., Horáček and Bosák 1989; Fig. 4).

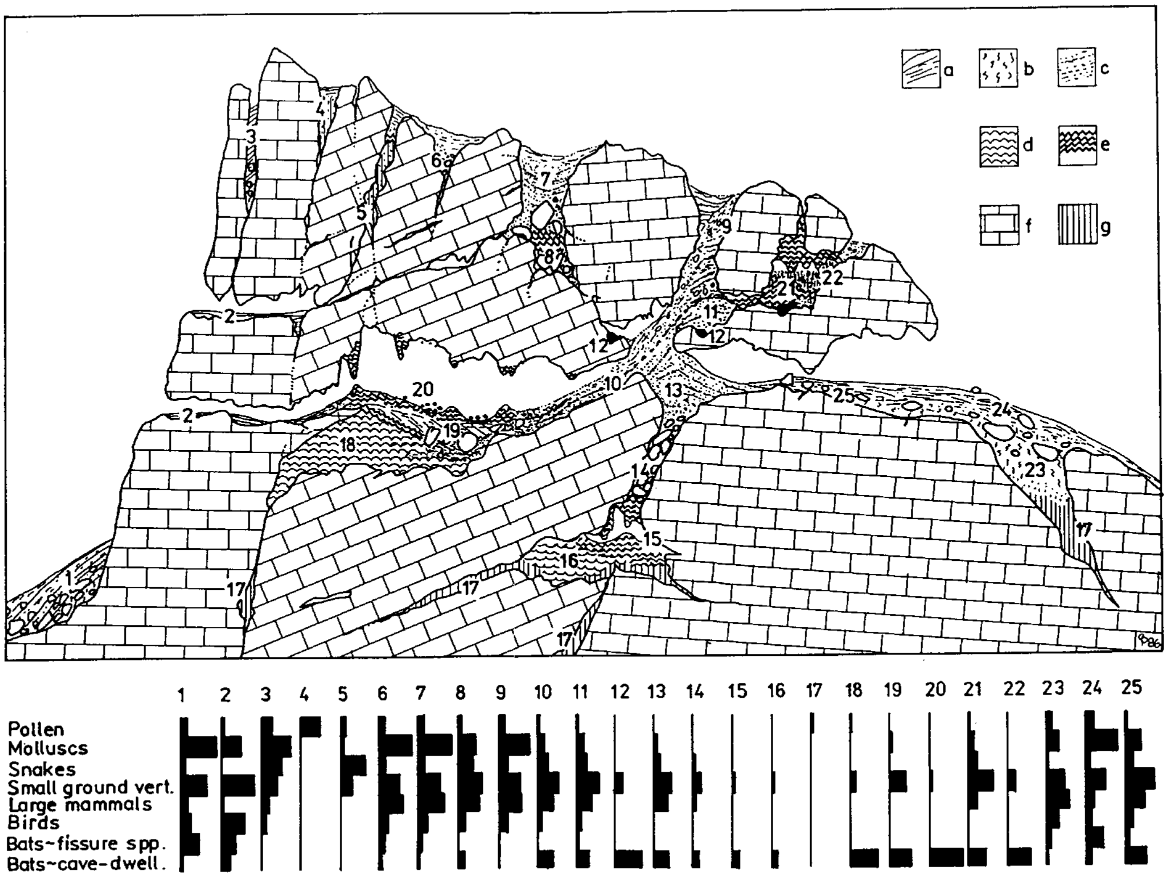

Fig. 4. A sketch of common types of karst infills and their fossil content. Note the appearance of remains of ancient fills of the inner-cave facies (12) preserved in wall niches, which may lie in the direct contact with much younger deposits (11) or those preserved in different but neighboring cavities (21 vs. 11). Collapse of sedimentary plugs and redeposition may also occur in caves (10), which may also cause serious confusion unless detailed lithological studies are done (see e.g., situation on sites 19 and 10). A – Holocene soils and related deposits, b – loess base of Holocene deposition, c – sequence of Pleistocene and earlier surface deposits, d – former infill of the inner-cave facies, frequently fluvial, e – flowstones, f – carbonate rock, g – ancient residua of strongly weathered surface or subsurface sediments, mosty non-calcareous (from Kordos and Horáček 1989, with permission).

Evaluation of dating results of karst records depends, as in other geologic records, on uncertainties, which vary with the geologic context, age range, and methods applied (Sowers and Noller 2000, p. 8-9). According to these authors, sources of uncertainty can be found in: (1) analytical error; (2) natural variability in sample quality and suitability; (3) geologic context errors; (4) calibration errors, and (5) violations of assumptions.

The best reviews of dating methods are offered by Geyh and Schleicher (1990), Noller, Sowers and Lettis (Eds., 2000), and Bradley (1999); some useful data can be found also in Faure (2001).

Numerical-ages

Numerical-ages are generally subdivided to isotopic, radiometric and sidereal (Colman and Pierce 2000, p. 3). Geyh and Schleicher (1990) divided only the radiometric methods, recognising those using (1) parent/daughter isotope ratios; (2) dating based on radioactive disequilibrium of the U, Th, and Pa decay series, and (3) age determinations using radiation damage. Methods (1) and (2) of the Geyh and Schleicher (1990) classification correspond to isotopic methods of Colman and Pierce (2000), and method (3) is the equivalent of radiometric methods. The U-Pb method was recently applied to about 92 Ma old spar fill in paleokarst in Guadelupe Mts., U.S.A. by Lundberg, Ford and Hill (2001).

TABLE 3

Review of isotopic dating methods I - parent/daughter isotope ratios

Dating method	Dating range	Suitable materials
$^{138}La/^{138}Ce$	>Ga	Basic rocks, acid rocks, pegmatites
$^{138}La/^{138}Ba$	>Ga	REE-bearing minerals
$^{207}Pb/^{206}Pb$	>Ga	igneous, metamorphic rocks, sulfides
$^{176}Lu/^{176}Hf$	>500 Ma	REE-bearing minerals
$^{187}Re/^{187}Os$	>200 Ma	meteorites, molybdenite, ultrabasic magnatic rocks
U/Xe_{sf}	U-minerals > 100 Ma terrestrial rocks >1Ga	U-bearing minerals; terrestrial rocks
Xe_{sf}/Xe_n	> 100 Ma	U-bearing minerals
$^{40}K/^{40}Ca$	> 60 Ma	high K content, low Ca content (K/Ca>50) - lepidotite, muscovite, biotite, K-fedspars, salt minerals
$^{147}Sm/^{143}Nd$	> ca 50 Ma	old, especially basic igneous rocks, high grade metamorphics, whole-rock and mineral samples, great resistance of the system
$^{87}Rb/^{87}Sr$	> 10 Ma	minerals and whole-rock samples, magmatic and metamorphic rocks, sediments with limitations (authigenic clay minerals) salt minerals - problems low temperature of metamorphism
Kr_{sf}/Kr_n	> 10 Ma	U-bearing minerals
$^{129}Xe/^{136}Xe$	5-100 Ma	U-bearing minerals
Common Lead Method	> Ma to Ga	Pb-bearing minerals with low or no U content, whole-rock (igneous)
$^{238}U/^{206}Pb$ $^{235}U/^{207}Pb$ $^{232}Th/^{208}Pb$	< 0.1 - > 100 Ma	U- and Th-bearing minerals in igneous and metamorphic rocks (esp. zircon and monazite), U-bearing opal and paleokarst calcite
$^{40}K/^{40}Ar$	>100 ka (K-feldspars) >3-5 Ma (alunite, jarosite)	K-bearing minerals from igneous, metamorphic and sedimentary rocks feldspars, mica, amphibole, glauconite, clay minerals, whole-rock materials - volcanic rocks, particularly basalts
$^{39}Ar/^{40}Ar$	ka-4.5 Ga	K-bearing minerals from igneous and metamorphic rocks with low Ca content (mica, alunite, amphibole), sedimentary rocks suitable sometimes (glauconite, clay minerals), K-bearing sulfides

Note: The table was compiled according to data in Geyh and Schleicher (1990); Noller, Sowers and Lettis (Eds. 2000); Faure (2001); White (1988), and Ford and Williams (1989). Some data were kindly provided by H. Hercman (Warsaw, Poland).

Isotopic methods

Isotopic methods measure changes in isotopic composition due to radioactive decay and/or growth (Colman and Pierce 2000). The *methods of parent/daughter isotope ratios* (Table 3) are based on radioactive decay: for each parent atom that decays, a stable daughter isotope is formed, either directly or as the end product of a decay series (Geyh and Schleicher 1990, p. 51). The number of decays depends on the quantity of parent nuclides. The decay of each radionuclide is characterised by (1) the kind(s) of radiation they emit (alpha, beta, spontaneous fission, beta-plus decay and orbital electron capture), (2) the energy(ies), and (3) the half-life (Geyh and Schleicher 1990, p. 25). Various radioactive isotopes have different half-lives ranging from several years (^{210}Pb) to billion of years (^{187}Re). This makes geochronological studies possible over the entire range of possible ages. The methods are based on long-lived radionuclides, therefore the application to Quaternary studies is almost excluded (Geyh and Schleicher 1990, p. 53).

The *method of dating with cosmogenic radionuclides* (Table 4) is based on nuclear reaction of cosmic rays with gas molecules in the stratosphere and troposhere producing many radionuclides. Samples must have existed in closed system conditions since the beginning of the aging period, i.e. since the geochronological clock was reset to zero (Geyh and Schleicher 1990, p. 158). Most methods are based on, first, the insolation of the material and then its burial at depths too great for cosmic ray penetration (e.g. in most caves or karst deposits).

The *methods of radioactive disequilibrium of the U, Th, and Pa decay series* are based on radioactive disequilibrium utilising the time-dependence of geochemical disturbances of the radioactive equilibrium between parent and daughter isotopes of the natural radioactive decay series of ^{238}U, ^{235}U and ^{232}Th, whose end members are stable lead isotopes (Ivanovich and Harmon, Eds. 1992; Geyh and Schleicher 1990, p. 213).

TABLE 4

Review of isotopic dating methods II - cosmogenic radionuclides

Dating method	**Dating range**	**Suitable materials**
^{129}I	3-80 Ma	buried organic matter and its derivatives
^{53}Mn	1-10 Ma	meteorites, ice and pelagic sediments
^{26}Al/^{10}Be	0.1-10 Ma	ice, marine and lacustrine sediments, corals, organic matter, manganese nodules
^{81}Kr	0.05-10 Ma	groundwater and ice
^{26}Al	0.1 - 5 Ma	ice, pelagic sediments, manganese nodules
^{36}Cl	0.1-3 Ma	old groundwater, soils, ice, glacial materials
^{10}Be	0.01-15 Ma	carbonate-free pelagic sediments, ice, manganese nodules, quartz pebbles
^{10}Be/^{36}Cl	X0-X00 ka	ice
^{41} Ca	20-400 ka	bones, secondary carbonates
^{14}C	0.3-30 (55) ka	organic matter, peat, humus, bones, tissues, carbonate shells, corals, travertines, speleothems, soils, groundwater, ice
^{39}Ar	0.1-2 ka	Ice
^{32}Si	0.1-1.5 ka	marine siliceous materials
^{3}H	< 100 a	Groundwater
^{3}H/^{3}He ^{3}He	< 100 a	Ice
^{22}Na	1-30 a	shallow groundwater

Note: The table was compiled according to data in Geyh & Schleicher (1990), and Noller, Sowers & Lettis (Eds. 2000). Some data were kindly provided by H. Hercman (Warsaw, Poland).

TABLE 5

Review of isotopic dating methods III - radioactive disequilibrium of the U, Th, Protactinium decay series

Dating method	**Dating range**	**Suitable materials**
U/He	< 30 Ma	non-recrystallised aragonite (marine fossils, corals)
$^{234}U/^{238}U$	50 ka – 1.5 Ma	marine molluscs, corals, lacustrine and pelagic sediments, speleothems
$^{230}Th/^{234}U$	< 100 a - 600 ka	fossils, bones, travertines, speleothems, oolite, manganese nodules, marine phosphorites, marine hydrothermal deposits
$^{230}Th_{excess}/^{232}Th$ $^{230}Th/^{238}U$	- 300 ka < 1 Ma	marine carbonates, manganese nodules, glass shards (volcanic ash), fish bones+teeth, lacustrine sediments with clay minerals igneous rocks phosphorite deposits
$^{230Th}{}_{excess}$	< 300 ka	deep sea sediments, manganese nodules
$^{231}Pa/^{235}U$	0.1-200 ka	fossils, bones, oolite, manganese nodules, marine phosphorites, less often travertines, speleothems; U-content several ppm
$^{231}Pa/^{230Th}$	0.1-200 ka	U-rich marine carbonate (corals mollusc shells)
^{226}Ra	< 200 ka	marine sediments, ice
$^{231}Pa_{excess}/^{230}Th_{excess}$	< 150 ka	pelagic sediments
$^{231}Pa_{excess}$	< 150 ka	pelagic sediments, corals, manganese nodules
^{210}Pb	< 150 a	lacustrine, fluvial and coastal marine sediments, coral, peat, ice
^{224}Ra ^{228}Ra	< 100 a	corals, Fe-Mn nodules in lakes
$^{228Th}{}_{excess}/^{232}Th$	< 10 a	High rate deposition in lakes, deltas, estuaries, along coast
Ra/Rn	30 – 100 days	groundwater residence time
$^{234Th}{}_{excess}$	< 100 days	short-term reworking and diagenesis

Note: The table was compiled according to data in Geyh and Schleicher (1990); Noller, Sowers and Lettis (Eds. 2000); White (1988), and Ford and Williams (1989). Some data were kindly provided by H.Hercman (Warsaw, Poland).

The principle of all isotopic methods is that the system has to be closed after deposition, only under such conditions can radioactive equilibrium be gradually established. It means that any disturbance occurring during the evolution of the equilibrium (starting with the closure of the system) can lead to the stopping or resetting of the radiometric clocks. The nature of the „disturbance“ depends on the sensitivity of the system, which mostly closes during the crystallisation of rock-forming minerals from magmas or solutions. The geochronometer can be stopped by heating, recrystallisation, diagenetic processes such as leaching, or corrosion leading to opening of the system or adjustment to new conditions (e.g., heating/cooling).

The review of isotopic methods is given in Tables 3-5 summarising only principal data of each method (dating range and suitable materials).

Radiometric methods

The methods are based on the interaction of non-conducting solids with ionising alpha, beta, gamma, and cosmic radiation that changes their physical and chemical properties (e.g., deffects in crystal lattice). The changes are known as radiation damage. The age determinations are based on two types of damage: (1) electron shell phenomena, and (2) lattice phenomena (Geyh and Schleicher 1990, p. 253-255). The review of methods is given in Table 6.

Fission-track method is a radiogenic method of age estimation based on accumulation of damage trails left by nuclei that are expelled during fission decays of ^{238}U. The method can be applied to minerals with relatively high U content (e.g., apatite, zircon, sphene, volcanic glass). It can be used for direct age determination and for indirect date estimates. Tracks in apatite are partially or entirely erased by increased temperature (110-135 ^{o}C), which corresponds to a depth of 3-6 km at normal geothermal gradient. This behaviour has been utilised for dating of unroofing, as lesser heat causes reduction in fission-track ages and reduction of fission tracks (Dumitru 2000).

Thermoluminescence methods are based on lattice defects in common minerals (e.g., quartz, feldspars) formed during crystallisation or from exposure to nuclear radiation. Heating of sediments causes vibration of mineral lattice and eviction of timer-stored electrons from traps (Forman, Pierson and Lepper 2000). Geyh and Schleicher (1990, p. 257) cite different age ranges for different materials and there can be numerous errors resulting from different sources for the materials and their exposure (see review in Forman, Pierson and Lepper 2000).

Electron spin resonance is based on lift of electrons by ionising radiation from the valence band to a conduction band. Some electrons fall into quasi-stable traps at "forbidden" energy levels. Traps occupied by a single electron act as paramagnetic centres, whose density can be measured by ESR (Geyh and Schleicher 1990, p.273).

Numerical-ages are provided also by numerous other methods, which have been applied especially in Cenozoic geochronology (see in Noller, Sowers and Lettis, Eds. 2000). *Dendrochronology* is based on variations in annual growth rings of trees. There are records extending back more than 7 ka. *Varve dating* in laminated sediments is based on annual depositional cycles, especially in lakes. The method can be applied for sediments 18 ka old, i.e. deposited since the last glacial maximum. *Sclerochronology* is the measurement or estimation of ages or time intervals from the growth patterns or inclusions contained in the mineralised biogenic deposits of animals and plants. The method has been applied on corals, molluscs, fish otoliths. Historical records are useful for dating historical events (e.g., collapses, earthquakes).

Calibrated-ages, relative-ages

Calibrated-age methods can provide approximate numerical ages. Relative-age methods provide an age sequence and most also provide some indication of the magnitude of age differences between the members in a sequence (Colman and Pierce 2000, p. 4). The methods of this type are specially chemical and biological methods and geomorphic ones.

TABLE 6

Review of radiogenic dating methods

Dating method	**Dating range**	**Suitable materials**
Fission track	20 ka-2.7 Ga	Direct age -minerals, obsidian, glass (natural and man-made), tectites, petrified wood, etc.) Indirect - age of cooling of some minerals - uplift and erosional history
Thermoluminescence	< 500 ka	archeological objects, quartz and feldspars, flint tools, shells, bones teeth, polymineral fine-grained samples, lava (plagioclase), tectites, volcanic glas, loess, travertine and speleothems, fossil calcite shells
Optically simulated luminescence	1-700 ka	eolians, fluvial, glacial sediments, quartz, zircon
Electron spin resonance (ESR and EPR)	25-50 ka to > 1 Ma (?100 Ma)	fossils, speleothems, travertine, caliche and vein fillings, pelagic sediments, ceramics, cooling ages of quartz, feldspars, silicates, glass, apaptite etc., crystallization age of gypsum
Exo-electron method (TSEE)	< 100 ka	bones and dentin
Thermally simulated current (TSC)	1-2 Ma	basalts only
Alpha-recoil track	> 100 ka	crystallisation of rock-forming minerals (esp. mica), ages of bones and dentin, low U content

Note: The table was compiled according to data in Geyh & Schleicher (1990); Noller, Sowers & Lettis (Eds. 2000); White (1988), and Ford & Williams (1989). Some data were kindly provided by H. Hercman (Warsaw, Poland).

TABLE 7

Review of chemical and biological methods

Dating method	Dating range	Suitable materials
Amino-acid racemization	< 500 ka theoretical range < 5 Ma	dating fossils matter that contains amino acids:bones, teeth, foraminifera, coprolites, mulluscs, land snails, marine phosphorites, tuffs, carbonate mud and oolite, speleothems, wood
Amino-acid degradation	up to Miocene	molluscs, foraminifers
Obsidian hydration	0.01->1 Ma	obsidian, ignimbrite, basaltic glass, fused shale, slag, vitrophyre, other natural glasses
N and collagen dating bones	< 100 ka	skeletal materials
F and U dating bones	up to Pliocene	skeletal materials

Note: The table was compiled according to data in Geyh & Schleicher (1990); Noller, Sowers & Lettis (Eds. 2000); White (1988), and Ford & Williams (1989).

Chemical and biological methods are based on the assumption that certain reaction rates (e.g., diffusion, exchange, oxidation, hydration) are at least nearly constant. The age is estimated from the initial and end concentrations of suitable reactants or products (Geyh and Schleicher 1990, p. 345). The *amino-acid racemization method* is based on the slow conversion of amino acids after an organism has died. The *amino-acid degradation method* is based on the natural degradation (mainly dehydration) of the ABA acid (Geyh and Schleicher 1990, p. 355). In the *obsidian hydration method* glasses adsorb water on the surface, where it becomes chemically bound, forming a hydrated layer. The process is diffusion controlled, so the layer grows very slowly. The diffusion front of the hydrated layer is a sharp boundary (Geyh and Schleicher 1990, p. 362). Dating of bones by the *nitrogen and collagen method* is based on the rate of protein decomposition, which is influenced by numerous natural factors. The *fluorine-chlorine-apatite method* in combination with the collagen method was modified by Wyszocański-Minkowicz (1969) to relative dating of bones identifying climatic conditions of bone fossilisation. Nevertheless, the recent data indicate that the expectations are very far from the reality and the method does not function. The *fluorine or uranium methods* utilises the fact that skeletal remains continually take up F and U from groundwater via an irreversible ionic exchange. Both methods are very rough with low precision (Geyh and Schleicher 1990, p. 356-357 and 336-370). There are also other methods, like rock-varnish method, lichenometry, soil chemistry applied in Quaternary geochronology (see in Noller, Sowers and Lettis, Eds. 2000, p. 241-292) or chemical electron-spin-resonance dating, molecular (protein and DNA) clocks, Ca diffusion and cation-ratio methods (see in Geyh and Schleicher 1990, p. 359-369).

Correlated-ages

Correlated-ages are based on the methods of classical geology, geochemistry, geophysics, paleontology and, archeology, e.g., paleontology and stratigraphy, paleomagnetism and magnetostratigraphy, climatic correlations and stable isotope studies, astronomical correlations, tephrochronology, archeology. Principles of these methods are summarised in various textbooks (e.g., for Quaternary in Noller, Sowers and Lettis, Eds. 2000). Combinations of methods have been often applied, e.g., paleontology/stratigraphy with magnetostratigraphy or stable isotope studies or astronomical variations. Particularly useful is the combination of correlated-age methods with numerical-age determination of some marker horizons.

The methodology applied to obtain correlated-age results depend on the nature of the geologic material filling the (paleo)karst and on the types of karst. The fills of exokarst landforms such as sinkholes offer more possibilities for the preservation of fossil fauna and flora than do cave interiors. Troglobitic fauna and flora are usually much too small in number and volume to be significant (Ford and Williams 1989, p. 367). Therefore, fossil remains within a cave that come from the surface (carried in by sinking rivers) or from trogloxenes (e.g., bats, some birds, some mammals) are more important. Airborne grains (pollen, volcanic ash) can only be important when favourable air-circulation patterns are developed within a cave.

There are also numerous *geomorphic methods*, applied especially to young – Cenozoic – landscape and coast evolution. Methods are in general summarised in Noller, Sowers and Lettis (Eds., 2000).

Stratigraphy. The duration of stratigraphic unconformity can be determined by its chronostratigraphic representation (Esteban 1991, p. 92) based on (1) minimum gap (the time interval not represented by the sedimentary record in the area, caused either by complete erosional removal or by nondeposition. The minimum gap corresponds to the difference between the youngest age of the truncated section and the oldest age of the onlapping section), and (2) maximum gap (the maximum time interval absent in the sedimentary record in the area. The maximum gap corresponds to the difference between the age of the truncated section and the age of the youngest bed of the onlapping section; Fig.5).

The stratigraphic order in sedimentary sequences is governed by the *law of superposition*, according to which under normal tectonic settings the overlying bed is younger than the underlying one. The law is valid for the majority of sedimentary sequences. However, the karst environment represents one exception. Owing to the dynamic nature of karst, its polycyclic and polygenetic character, karst records can be damaged by the simple process of redeposition. In several places in the Czech Karst (Czech Republic) during the Early Quaternary (Biharian stage), destruction of the roofs of some caves and re-opening of fossilised vertical shafts (drawdown vadose connections) caused the excavation of pre-Quaternary fossil-bearing sediments and their deposition into younger caves. In Koněprusy Caves, such re-deposited fill from a vertical chimney was washed into a block collapse in the form of pseudo-matrix (see Bosák, Horáček and Panoš 1989). Contamination of younger deposits by re-deposited fossil-bearing sediments has been known elsewhere in caves (e.g., re-deposition of Cretaceous forams in Pleistocene deposits in the Moravian Karst). In Castleguard Cave (Canadian Rocky Mts.) there are Cretaceous pollen in basal varve layers of Würmian age (D.C. Ford, pers. comm. 2002). Well-known are also sandwich structures, described by Osborne (1998). Younger beds are inserted into voids in older ones. Those processes degrade the record in karst conservers (Horáček and Bosák 1989).

Biostratigraphy. Reinforcing the law of superposition are the use of *index fossils* (a widely distributed fossil that occurs only in one stratigraphic horizon), and the concept of *facies* (different conditions can at one and the same time create different assemblages at different sites, while almost identical assemblages may derive from different time periods). Biostratigraphy is based on vertical subdivision of geologic time according to fossil fauna and flora, which dominated at the studied time. Biostratigraphic systems may be defined either as a *range zone*, i.e. by means of the first and the last appearance dates of suitable index forms, or as an *assemblage zone* if based on specific characteristics of community structure. The time interval of individual biozones depends on the general evolution velocities of living organisms, therefore intervals shorter than 0.3 Ma can scarcely be recorded by biozonation and the common resolution is 0.7 Ma (Jindrich Hladil, pers. comm. 2002). A useful correlation is given in Haq, Hardebol and Vail (1988) and Berggren et al. (1995) indicating that the resolution of individual biozones of different kinds of fossils range from more than 6.5 to about 0.3 Ma. For biostratigraphic zonation, the application of fauna/flora evolution differs for marine and terrestrial records: nevertheless the principles of zonation in marine and lacustrine sediments are very similar. Fauna and flora in the terrestrial domain are often facies dependent, influenced especially by climate. In the Cenozoic, mammalian biozones (MQ, MN) differ in duration in different regions as a consequence of migration velocities and routes (see Horáček and Kordos 1989). There is also known "mixing" of flora of Carboniferous and Permian affinities, e.g., in Czech Upper Paleozoic limnic basins; arid facies contain Permian flora deeply below Carboniferous/Permian boundary.

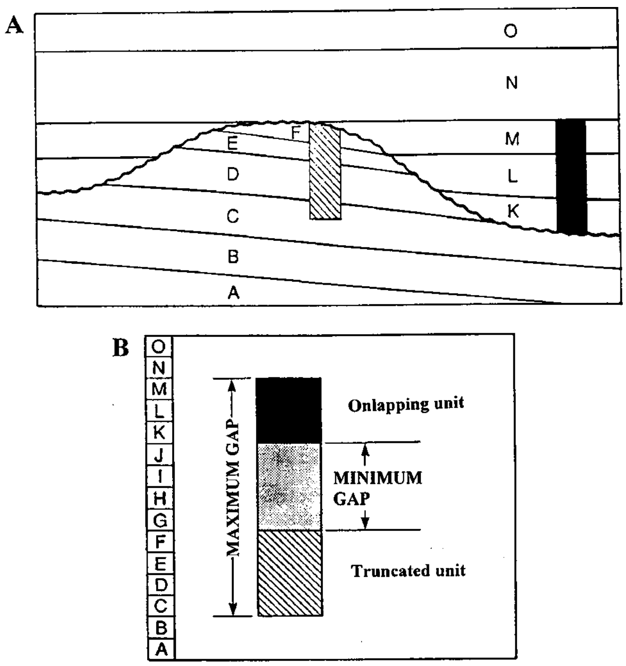

Fig. 5. Chronostratigraphic representation of an unconformity (modified after Esteban 1991).

Submerged caves may be characterised by peculiar biotopes containing very old elements with close ties to deep-sea fauna (e.g., recently in the Caribbean area). Caves can serve as refuges over very long time-spans, with highly conservative faunal assemblages. Such situations need to be recognised during the biostratigraphic interpretation of marine organisms found in cave facies, especially when studying transgressive tracts on karst surfaces (see also Horáček and Kordos 1989, p. 610).

Paleomagnetism and magnetostratigraphy. The method is based on variations in the polar declination, inclination and intensity of the Earth´s magnetic field. The changes are recorded in rocks by the orientation of magnetic minerals during their deposition or crystallisation. Use of records of ancient variations as a dating tool relies on matching the curves of declination and inclination in a given deposit with established curves (standard time-scales) that have been dated by independent methods (e.g., Ford and Williams 1989). The method faces numerous constraints and thresholds, especially where there is no independent dating of deposits by numerical-ages.

The complete reversals (excursions) of the field occur at 10^5-10^6 years and establish the principal time units (chrons). Nevertheless, there were periods when the polarity was stable for very long times, e.g., in the Cretaceous from about 107 to about 83 Ma (see e.g., Haq, Hardebol and Vail 1988). Most normal- or reverse-polarised deposits contain short-lived changes of polarity (subchrons) with durations from 10^0 to 10^2 ka. The combination of detailed micropaleontology with dense sampling for paleomagnetic analysis can result in high-resolution scales, e.g., a precision of about 5 ka on the Jurassic/Cretaceous boundary in the Tethyan realm (Houša et al. 1999) or even better for reversals in Pleistocene record combining paleomagnetism and thermoluminescence dating (Zhu and Tschu, Eds. 2001).

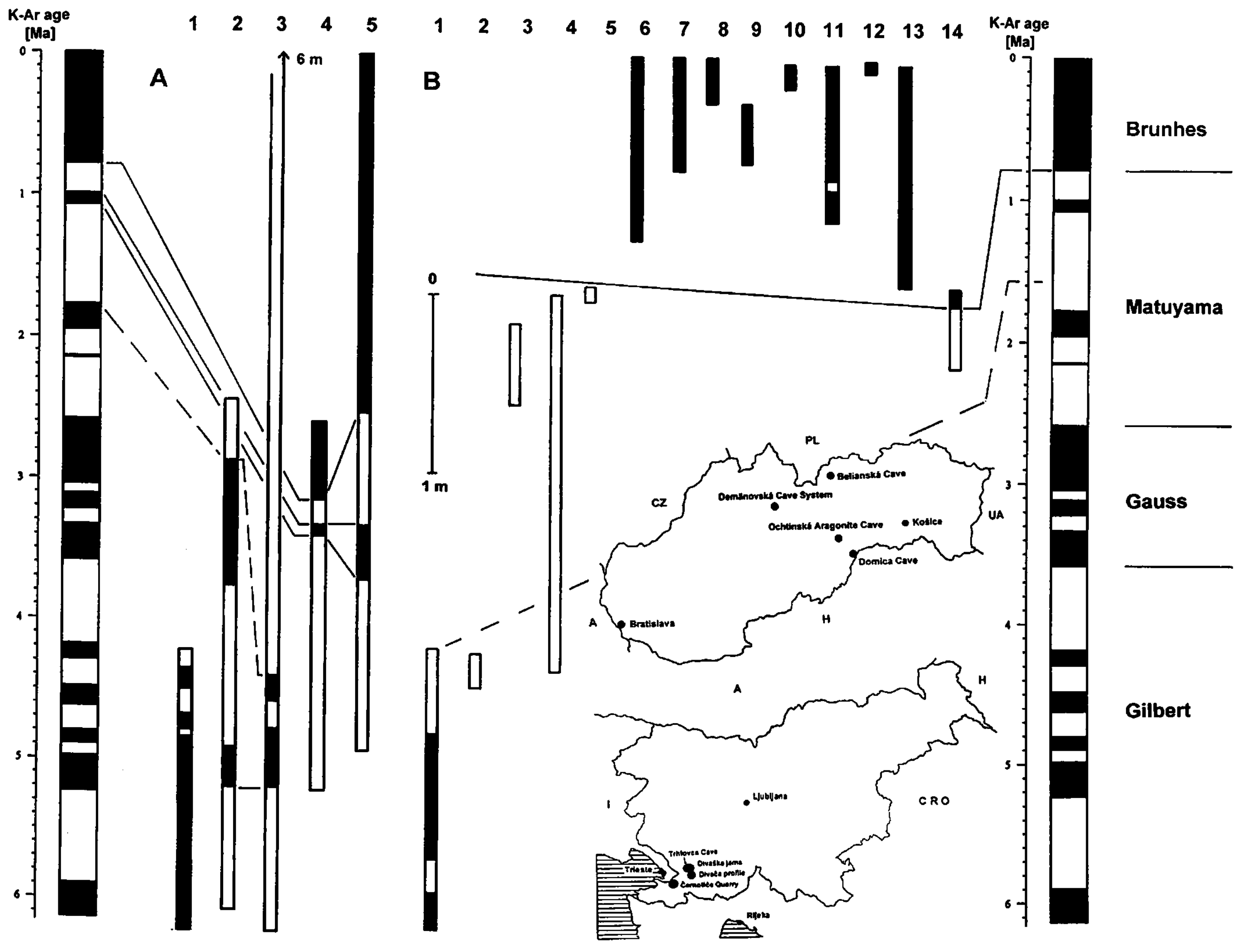

Fig. 6. Measured magnetostratigraphic profiles in some of Slovenian (A) and Slovak caves (B) and their correlation with the magnetostratigraphic chart of Cande & Kent (1995; after Pruner & Bosák 2001). A. Slovenia: 1 – Črni Kal-Černotiče, 2 – Kozina profile, 3 – Divača profile, 4 – Divaška Jama, 5 –Trhlovca Cave; B. Slovakia 1-2 – Belianská Cave, 3-7 – Demänovská jaskyňa Slobody, 8-9 – Demänovská jaskyňa Mieru, 10-13 – Domica Cave, 14 –Ochtinská Aragonite Cave.

The application of the method for dating clastic cave sediments has been limited by the complex conditions underground, i.e. it is often necessary to combine it with other methods offering numerical- relative- or correlate-ages. Moreover, the character of cave deposition results in numerous breaks in deposition, in which substantial timespans can be lost (Bosák et al. 2000; Pruner and Bosák 2001). The example of correlation of magnetostratigraphy results from selected caves in Slovakia and Slovenia is presented in Fig. 6. Paleomagnetic and magnetostratigraphic studies have been successfully applied also on calcite speleothems (e.g., Latham, Schwartz and Ford 1979, 1986).

The secular variations are quasi-periodic changes in declination and to lesser extent also of inclination. The changes are of smaller magnitude than those described as excursions and appear to be merely regional in extent. They presumably result from changes in the non-dipole component of the magnetic field. If the changes are dated independently, they can be used in chronostratigraphic time scales (Bradley 1999). The study of magnetosusceptibility of different age periods when adjusted to numerical- or correlate-ages represents also a useful tool for correlation or dating. The method can be used in deep-sea sediments, carbonate platforms, loess accumulations, etc. The content of ferro- and paramagnetic minerals is studied. Their contents are fixed during deposition and/or early diagenesis. Magnetosusceptibility stratigraphy has been applied to some Devonian carbonate sequences (Crick et al. 1997; Hladil et al. 2002) or for some Quaternary deposits (Kadlec et al. 2001). The changes in magnetosusceptibility are believed to be influenced by climatic conditions (temperature, humidity, winds) and, maybe, by Milankovich cycles.

Astronomical correlations. Orbital perturbations, known also as Milankovich cycles, reflect the astronomical cycles: the precession of the equinoxes (with a periodicity of 19 and 23 ka), obliquity of the ecliptic (41 ka) and eccentricity of the orbit (100 ka). It is widely believed that the orbital-forcing, Milankovich-rhythm mechanism is responsible for continental icesheet build up and the consequent sea-level changes, which can be recorded e.g., in Caribbean-model shallow marine carbonate sequences as erosional/karst surfaces and meteoric diagenetic changes (e.g., Tucker and Wright 1990). Astronomical cycles are well preserved both in marine and continental deposits, especially in laminated sequences and profiles with cyclic patterns. Most of studies indicate cycles of about 20-23 ka, 40-41 ka, 100 ka and 400-405 ka, which can be mutually superimposed. The detailed study of cyclicity of sediments, i.e. calibration of sedimentary cycles, or other cyclic variations in the geological record, to computed time series of the quasi-periodic variations of the Earth´s orbit, can result in cyclostratigraphic sequential scales. When calibrated by numerical dating (e.g., Ar/Ar single grain) they can substantially contribute to the construction of geological time scales (e.g., Neogene astrochronology in the Mediterranean; Krijgsman et al. 2002; Abdul Aziz et al. 2002) and to the improvement of previous models, e.g., the standard geomagnetic polarity timescale of Cande and Kent (1995) for the Cenozoic was age-corrected by astrochronology (Abdul Aziz et al. 2002).

Stable isotopic studies. Oxygen isotopic studies provide data to understand past environmental conditions, especially paleotemperatures. Relative abundance of oxygen ^{16}O and ^{18}O, the $^{18}O/^{16}O$ ratio, is compared with that in standards (PDB belemnite for solids and standard mean oceanic water – SMOW – for liquids; e.g., J.D. Wright 2000). If variations of marine stable isotope records are compared with numerical-ages and correlated-ages, a chronostratigraphic time scale can be constructed (Emiliani 1955; Shackleton and Opdyke 1973; Hays, Imbrie and Shackleton 1976; Imbrie et al. 1984). The oxygen isotope curve shows temperature changes influenced by glaciations. The time scale for the whole Quaternary has been established by this means. It is composed of 22 stages, with boundaries numerically dated by ^{14}C, K/Ar, Ar/Ar and U series dates and compared with paleomagnetic records and orbital variations. The stable isotope time scale has been often used for karst studies (e.g., Mylroie and Carew 1986; see also Ford and Williams 1989).

Correlation of cave levels and river terraces

Correlation of cave levels with river terraces has been relatively common in the past. Speleogenetic models of extensive areas were based on such correlation (e.g., the Czech Karst with the Berounka River, Czech Republic; Hromas 1968). Nevertheless, most of these correlations were limited to nearby rivers rather than to entire drainage areas (White 1988, p. 318). Sawicki (1909) defined so-called *evolution level*, i.e. connected with the piezometric surface and oriented towards the base level (see also Bögli 1981). This view allows to determined „cave levels“ even for deep phreatic or bathyphreatic systems, see Hölloch Cave System (Muotatal, Switzerland; Bögli 1966) with three main levels of bathyphreatic caves correlated with principal interglacials of the Alps.

White (1988) mentioned several examples of cave levels (water table *sensu* Ford 1968 or epiphreatic caves *sensu* Jennings 1985) correlated with the entrenchment of rivers, especially of the Mammoth Cave System (U.S.A.), where cave sediments showed good agreement between magnetostratigraphy and the model for its Tertiary-Quaternary evolution. Detailed analysis of factors influencing the interpretation of cave levels was summarised by Palmer (1984). Sharply defined cave levels with narrow vertical ranges (e.g., Mammoth Cave, Kentucky, USA) appear to have formed in response to intermittent episodes of rapid valley entrenchment, probably by headward erosion, followed by lengthy periods of virtually static base level.

Maybe the most conspicuous example of correlation of river terraces and cave levels has been in the Demänová Cave System (Demänovská Valley, Low Tatras Mts., Slovakia) developed by Droppa (1966) and mentioned in numerous textbooks (e.g., Bögli 1981, p. 116-119; Jennings 1985, p. 243-244). Droppa (1966) recognised 9 cave levels and correlated them to the well-developed terrace system of the Váh River. Recent detailed magnetostratigraphic (Pruner and Bosák 2001) and U-series dating of cave sediments and speleothems (Hercman et al. 1997) in the 4th and 5th cave levels (*sensu* Droppa 1966) has shown that the cave fill of these passages is older than the age of correlated terrace of the Váh River. From the combination of results, the 4th cave level was dry already at about 700 ka (the base of speleothem is ca 685 ka), although previous correlation with river terraces assumed the age of speleogenesis to Mindel 2, i.e. to ca 330-500 ka (Droppa 1972). Magnetostratigraphic data from higher cave levels from both Demänovská and parallel Jánská Valley indicate that the age of cave fill can be correlated with the age of sediments covering river terraces of the Váh River. Caves formed under phreatic and reworked under vadose conditions are therefore older. From the longitudinal sections of the cave system it is concluded that the evolution of passages followed the four state model of Ford (1968, 1971) and Ford and Ewers (1978). Upper levels represent rather deep phreatic caves with multiple deep loops later modified by vadose entrenchment and bypassing, while the lower cave levels can be correlated with nearly ideal watertable cave with minor shallow phreatic loops. Therefore, the cave levels should be correlated with the positions of respective karst springs rather than with terrace surfaces of the same or similar elevation, which can be lowered by subsequent erosion.

Conclusions

The precise dating of events during karst initiation, evolution and destruction is a highly risky task. Owing to the fact that karst and caves have been developing since the Archean, nearly all known dating methods can be applied. Paleokarst features can be fossilised by infilling and/or cover with a broad variety of rocks: marine and continental chemical and siliciclastic deposits, mineral deposits produced e.g., by weathering or hydrothermal activity, products of volcanism (lava, volcaniclastics). Recent karst surfaces and accessible caves can be covered/filled by a very similar spectrum of fills.

The methods determining the age of fills directly are based on physical, chemical and biological methods, plus methods of classical geology and stratigraphy. There are also indirect means of dating – correlation with correlative sediments not occurring in the karst itself. The range of age data produced by individual groups of methods substantially differs. There are geochronologic methods giving real dates – numerical-ages and ages based on correlation – calibrated (or relative)-age and correlate-age. The principal problem of dating of paleokarst features is in determining the duration of stratigraphic discontinuities. The longer are the discontinuities, the greater is the proportion of time not recorded in any correlated sediments (40 to 90 % of time can be missing in old platforms). Results of paleokarst evolution are best preserved directly beneath a cover of marine or continental sediments, i.e. under sediments, which terminated karstification periods or phases. The longer the stratigraphic gap the more problematic is precise dating of the age of the paleokarst, if it cannot be chronostratigraphically proven. Therefore, ages of paleokarsts have been associated chiefly with periods just or shortly before the termination of the stratigraphic gap. The characteristic time scale for the development of a karst surface landform or a conduit is 10 to 100 ka (White 1988, p. 304).

The dating of cave initiation and evolution, i.e. the origin of the void within the bedrock is more problematic. The age of the erosional cave falls between the age of the host rock and that of the oldest dated fill. With the inception theory, the true start of speleogenesis can hardly be estimated. Many caves contain only very young fills, older ones having been excavated during repeating cave exhumations/rejuvenations caused by changes in hydraulic conditions, spring position, climate, etc. The minimum age for the cave initiation phase is estimated to be a minimum of 10 ka and cave enlargement up to accessible diameters usually takes about 10-100 ka under favourable conditions.

The end of karstification occurs at the moment when host-karst rock together with its karst phenomena is completely eroded/denuded - the end of the karst cycle. In such case, nothing can be dated, all has been denuded. Karst forms of individual evolutionary stages (cycles) can be destroyed by erosion, denudation and abrasion, complete filling of epikarst and covering of karst surface by impermeable sediments, without the necessity of destroying the entire sequence of karst rocks (the cycle of erosion). Temporary and/or final interruption of karstification is caused by the fossilisation of karst due to loss of the hydrological function of the karst. Nevertheless, in contrast to living organisms, the development of the karst system can be „frozen“ and rejuvenated even for a multiplicity of times (polycyclic and polygenetic nature of karst). Further, the dynamic nature of karst can cause redeposition and reworking of classical stratigraphic order, making the karst record unreadable and problematic for interpretation.

The final answer to the question posed in the introduction is: according to my long-lasting experience, yes we can date karst processes and events; under extremely favourable conditions we can date the products of some processes very precisely by numerical dating and/or a combination of methods, but in a majority of cases we have to handle a number of unknown factors. To solve the problem we apply complex approaches, including geopoetry, more or less successfully depending on talent of student of the karst.

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