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Karst processes and time

Procesy krasowe a czas

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Abstract

Karst evolution depends particularly on 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 directly influence the intensity and extent of karstification. Unconformities influence the stratigraphy of the karst through the time-span that is available for subaerial processes. The end of karstification can also be viewed from various perspectives. The definite end occurs at the moment when the host rock, together with its karst phenomena, has completely been eroded/denuded. 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 succession of karst rocks. Temporary and/or final interruption of the karstification process can be caused by the "fossilisation" of the existing karst phenomena due to loss of hydrological activity. The shorter the time available for karstification, the greater is the likelihood that karst phenomena are preserved in the stratigraphic record. While products of short-lived karstification on shallow carbonate platforms can be preserved by deposition during a immediately succeeding sea-level rise, products of more pronounced karstification can be destroyed by various geomorphological processes. The longer the duration of subaerial exposure, the more complex these geomorphological agents are.

Keywords: karst, speleogenesis, geochronology, unconformities

Streszczenie

Rozwój procesów krasowych jest funkcją czasu oraz geograficznych i geologicznych warunków odsłonięcia skał. Im dłuższy czas ekspozycji skał na czynniki meteorologiczne, większy gradient hydrauliczny, większa ilość wody w układzie krasowym, tym bardziej zaawansowana jest ewolucja krasu. Intensywność i zasięg krasowienia zależą też od niezgodności stratygraficznych, czyli przerw w sedymentacji. Zakończenie rozwoju procesów krasowych rozpatrywać można w różnych kategoriach. Za definitywny koniec należy uznać czas, gdy skały podlegające krasowieniu ulegną całkowitej denudacji/erozji. O wiele częściej bywa, że zniszczeniu ulegają tylko formy krasowe, natomiast niżej położone skały systemu krasowego pozostają zachowane. Okresowe lub całkowite przerwanie procesów krasowych może być spowodowane przez fosylizację systemu krasowego, która zachodzi w efekcie zaniku aktywności hydrologicznej. Taka fosylizacja może być spowodowana przez metamorfizm, transgresję morską, pogrzebanie osadami kontynentalnymi lub skałami wulkanicznymi, w wyniku np. ruchów tektonicznych, zmiany klimatu, itp. Im krótszy jest czas krasowienia, tym większe jest prawdopodobieństwo zachowania śladów procesów krasowych. I tak, produkty krótkookresowej karstyfikacji na płytkich, okresowo wynurzanych platformach węglanowych mogą ulegać łatwemu zachowaniu poprzez pogrzebanie osadami deponowanymi podczas podniesienia poziomu morza. Natomiast efekty długotrwałego krasowienia bywają często niszczone przez późniejsze degradacyjne procesy geomorfologiczne. Charakter tych ostatnich jest tym bardziej skomplikowany, im dłużej trwa subaeralna ekspozycja skrasowiałych skał.

Słowa kluczowe: kras, speleogeneza, geochronologia, niezgodności

Introduction

Each process needs time to start, evolve and end. The level of process evolution depends particularly on (1) the intensity of the process itself, and (2) the available time. Process intensity is related to the input of energy (the production of entropy), which reflects external (e.g., climate, geographical position) and internal factors and conditions of rock exposure (e.g., lithology, tectonics; *cf.* Eraso 1989; Ford & Williams 2007; Ford 2002). The more time is available, the higher is the hydraulic gradient and the larger is the quantity of solvent medium (water) entering the system, the more the karst evolves in all its types (exo- and endokarst). The time available for evolution of the process consequently represents an essential factor.

The Symposium "Time and Karst" (March 2007, Postojna, Slovenia), where the present contribution was presented in full detail, was organized relatively shortly after another Symposium with the title "Evolution of karst: from prekarst to cessation" (EVOKARST; Gabrovšek, Ed. 2002). The themes of both symposia partly overlapped. In a contribution that I presented at EVOKARST (Bosák 2002), I dealt with time-related aspects of karst evolution and the dating of karst processes from the beginning to the end, because dating of karst processes requires that some kind of an absolute or relative age framework is obtained for the karst phenomena dealt with. It is therefore unavoidable that the present contribution contains some similar phrases, tables and figures as my EVOKARST contribution, although some of them have been substantially modified. There are two reasons: the tables and figures concerned are illustrative, and the progress in knowledge since 2002 has not changed our ideas or points of view.

The present contribution deals mostly with karst in carbonate rocks. A short review of other karst lithologies is added.

Karst: a special geological feature

The life – start, development, cessation – of a karst system still poses substantial problems. In contrast to most living systems, the development of karst systems can be „frozen" (halted) and then rejuvenated, which happens often several times (*cf.* Bosák *et al.*, Eds. 1989), so that karst deposits represent a special kind of geological record (Bosák 2002). When karst is hydrologically decoupled from the contemporary hydrological system, it becomes paleokarst (Bosák 1981, 1989; Bosák *et al.* 1989; Ford & Williams 2007), independent of whether the karstification is halted definitely or only temporarily. The most common reasons for such interruptions or cessations are metamorphism, mineralisation, marine transgressions/ingressions, burial by continental deposits or volcanic products, tectonic movements (uplift, subsidence), climatic change (desertification, glaciation), etc. (for a review, see Bosák 1989). The introduction of new energy (hydraulic head) to the system may cause reactivation of the karstification. The most common reasons for reactivation are regression, deglaciation and uplift (for a review, see Osborne 2002). Multiple reactivations are result in polycyclicity of karst formation, which is a characteristic feature (e.g., Panoš 1964; Ford & Williams 2007; Wright 1991; Osborne 2002). The polygenetic nature of many karsts features that evolved during several different steps should be stressed, too (Ford & Williams 2007); these may take the form of, for instance, an overprint of cold karst processes on earlier deep-seated/hydrothermal products, which themselves followed meteoric early diagenesis (e.g., Bosák 1997) or the succession of other processes (a.o., Osborne 2000, 2002; Osborne *et al.* 2006).

The introduction of a time scale for karst evolution poses philosophical problems, principally regarding (1) the precise definition of the beginning of karstification, and (2) the modes of preservation of any karstification products, recognising that karst rocks are more easily soluble than other rock types under specific conditions that depend also on the lithology (limestone, dolomite, gypsum, anhydrite, rock salt, quartzite). Preservation of the karst record is important because karst areas preserver the geological and environmental past. This is of special importance for the terrestrial (continental) history, where correlative sediments are mostly missing, but also for the marine record (Horáček & Bosák 1989).

The karst environment facilitates both the preservation and the destruction of paleontological remains. On one hand, karst is well known for its wealth of paleontological sites (a.o., Horáček & Kordos 1989), on the other hand most cave infillings are completely sterile, especially the inner-cave facies. A problematic feature of karst records is that reactivation of processes may make the unravelling of the record impossible, for instance as a result of the mixing of karst infillings of different ages (due to collapse, redeposition, etc.: a.o., Horáček & Bosák 1989; Osborne 1998). 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 & Noller 2000; Bosák 2002).

The time frame of karst

In general, the kind of stratigraphic discontinuities, i.e. intervals of non-deposition (disconformities and unconformities: Esteban 1991), directly influences the intensity and extent of karstification. The higher is the order of a discontinuity under study, the bigger are the problems regarding the time frame of the processes and the dating of the successive events.

Unconformities

The beginning and the end of karst development is clearly associated with conformities, unconformities and disconformities. Esteban (1991), in an excellent review following a sequence-stratigraphic approach, outlined the role of non-depositional events (stratigraphic discontinuities) in karst evolution. Different ranks of stratigraphic discontinuity represent the various time gaps in deposition that have been available for dissolution (karstification; see also Moore 2001).

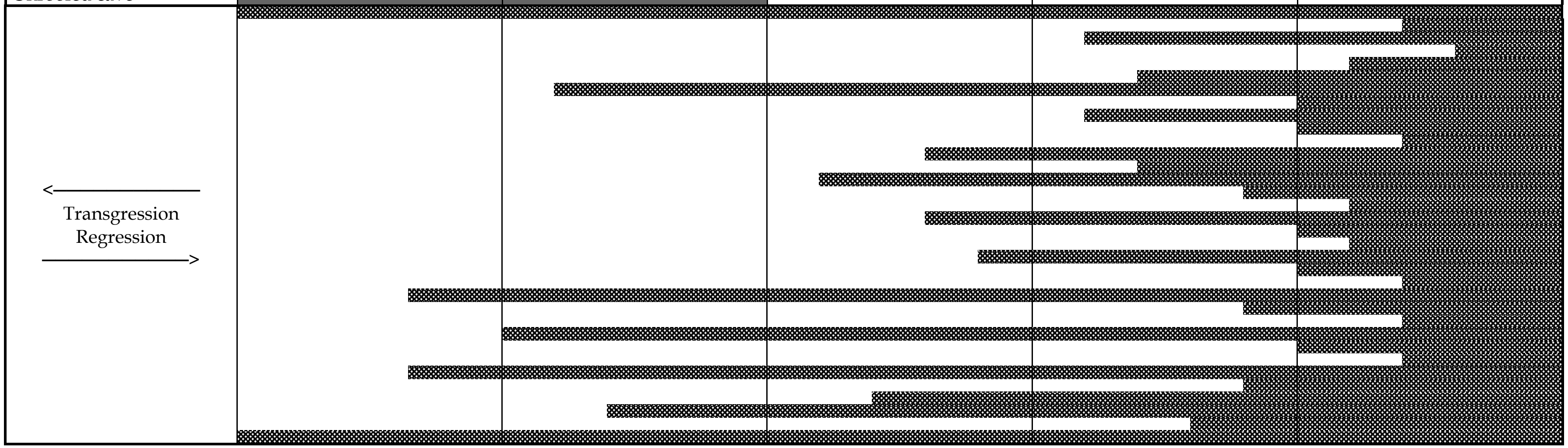
The stratigraphic discontinuity (gap) represents the chronostratigraphic interval(s) missing as a result of non-deposition (hiatus) and/or lithostratigraphic interval(s) missing due to erosion. Excluding conformities, Esteban (1991) proposed the classification of unconformities into single and composite types, both with measurable stratigraphic gaps (during which karst can be developed). Conformities have no measurable stratigraphic gap and correspond mostly to bedding planes (no karst development). 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). Most (paleo)karst features include composite unconformities, representing long time-spans without deposition.

The hierarchy of stratigraphic discontinuities in Tab. 1 is based on the original idea of Esteban (1991, Fig. 3.5) but expressed in time levels. This modification better illustrates the problem of stacking of unconformities, and clearly demonstrates that the lower is the unconformity order, the lower time is available for any subaerial process to act (karstification, weathering, erosion, denudation, deposition, etc.). On the other hand, the lower is the unconformity order, the better is the dating of a stratigraphic gap. The more time is available, the better-developed subaerial features can be expected. Longer periods of non-deposition are characterized by both the formation and the destruction of karst forms, especially in favourable paleoenvironmental and paleotectonic settings.

Tab. 1. Evolution of selected karst features in time against the background of a transgression/regression set within one hypothetical karst period related to unconformity order

Tab. 1. *Ewolucja zjawisk krasowych w trakcie jednego hipotetycznego okresu krasowego*

Feature/Order+	**1**	**2**	**3**	**4**	**5**
Unconformity+	Megaunconformity	Superunconformity	Regional unconfornity	Parasequence boundary	"Bedding plane"
*Carribean model**	Interregional karst		Local karst		Depositional karst
*General model***	Karst period		Karst phase Type 1		Karst phase Type 2
Geological setting	Craton/Platform – centre	Craton/Platform + margins	Depositional basin		
Time (Ma)	X00–X0	X0–X	X–0.X	0.X–0.0X	0.0X–0.00X
Freshwater lens					
Protosol					
Caliche					
Soil					
Weathering profile					
Karren					
Sinkhole					
Cave					
Cave system					
Hypogenic karst					
Hydrothermal k.					
Early karst*					
Mature karst*					
Buried karst**					
Rejuvenated k.**					
Relict karst**					
Unroofed cave§					



+ *sensu* Esteban (1991); * *sensu* Choquette & James (1988); ** *sensu* Bosák *et al.* (1989); § *sensu* Mihevc (1996). Weathering profile = more evolved weathering cover (like laterite, bauxite, kaoline, etc.). Hypogenic karst = deep-seated karst, interstratal karst, intrastratal karst, subjacent karst, subrosion

Terminy według: + *Estebana (1991);* * *Choquette'a & Jamesa (1988);* ** *Bosáka et al. (1989);* § *Mihevca (1996). Weathering profile = pokrywa zwietrzelinowa typu laterytów, boksytów, kaolinów. Hypogenic karst = kras podziemny, międzywarstwowy, wewnątrzwarstwowy, subrozja*

Stratigraphy of karst

The order of unconformities influences the stratigraphy of the karst due to the time involved in subaerial processes (Tab. 2). There are two general approaches to karst development (Esteban 1991):

(1) The *Caribbean model* is characterised by a short exposure time, unstable carbonate mineralogy, shallow burial, minor tectonics, a minor deep (fresh-water) phreatic zone with dominant primary and fabric-selective porosities, restriction to tropical to semi-arid environments, diffuse recharge-diffuse flow only, affected by mixing processes in the marine zone but not by hydrothermal mixing. However, geothermally-driven convection of groundwater has been detected in some Caribbean-type settings (Rougerie & Wauthy 1993).

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

The two karst-development models are also reflected in two general systems of the karst stratigraphy based on: (1) a carbonate sedimentological/sequence-stratigraphic approach (Choquette & James 1988), and (2) general karst models (Bosák *et al.* 1989).

Tab. 2. Stratigraphic discontinuities, time gaps, and stratigraphy of karst (modified after Bosák 2002)

Tab. 2. *Niezgodności i luki stratygraficzne a klasyfikacja stratygrafii procesów krasowych (wg Bosáka 2002, zmodyfikowane)*

STRATIGRAPHIC DISCONTINUITIES		ORDER	TIME GAP SCALE		CORRESPONDING STRATIGRAPHIC UNITS	STRATIGRAPHY OF KARST	
			Ma	Chrono-stratigr.		James & Choquette, Eds. 1988	Bosák *et al.*, Eds. 1989
UNCONFORMITIES COMPOSITE	**erathem uncorformity megauncorformity**	**1**	> 200 > 60	**erathem system series stage**	**megasequence supersequence set supersequence**	**inter-regional karst**	**karst period**
	superuncorformity set superuncorformity	**2**	30 4–12				
SINGLE	**regional uncorformity (sequence boundary)**	**3**	~ 1	**biozone**	**depositional sequence**	**local karst**	**Type 1 karst phase Type 2**
CONFORMITIES	**syntectonic uncorformity**	**3–4**	0.0X-1	variable			
	boundary of shoaling cycles	**4**	0.0X	not recognisable	**parasequence**	**depositional karst**	
	bedding plane	**5**	0.00X		**bed**		

Choquette & James (1988) distinguished the following three karst forms.

(i) *Depositional karst* forms as a natural consequence of sediment accretion at and around sea level. This is to be expected within the sediment packages that typify carbonate platforms. It is most commonly associated with meter-scale depositional cycles (Choquette & James 1988). Esteban (1991) stressed that the depositional karst of Choquette & James (1988), which is associated with parasequence boundaries, reflects a *Caribbean model* of karst development

(ii) *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 time-span involved, the effects of exposure can vary from minor to extensive with the development of exo- and endokarst (Choquette & James 1988).

(iii) *Interregional karst* is much more widespread. It is related to major eustatic-tectonic events, and results in karst terrains that may exhibit profound erosion, a wide variety of karst features, and deep, pervasive dissolution (Choquette & James 1988). They noted that in some cases it may be difficult to distinguish the products of local and interregional karsts. Esteban (1991) stressed that interregional karst results from complex evolution producing the composite unconformities karst, and represents the general model of karst.

Bosák *et al*. (1989) distinguished the following two intervals of karst formation.

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

(b) A k*arst period*, which they define as long-lasting times of groundwater circulation and continental weathering, terminated by an ensuing marine transgression. These periods are recognised by higher-order unconformities or disconformities (=*interregional karst* of Choquette & James 1988). The resulting 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. These 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 be confined to individual modern continents. These long periods display diachronicity and many less pronounced phases. They are longest in duration and most complex at mountain crests and become gradually shorter on the mountain slopes and their wide fringes along adjacent continents. These periods result from major changes in plate-motion patterns and they divide structural complexes corresponding to orogenic/geotectonic cycles (Głazek 1973) with durations of about 200–250 Ma or more.

Interregional (paleo)karst and products of karst periods can be linked with the composite unconformities karst of the 1st and 2nd orders *sensu* Esteban (1991). Such products can be correlated over extensive regions, as shown by the Paleozoic post-Sauk and post-Kaskaskia karsts (*cf*. Palmer & Palmer 1989 and case studies in James & Choquette, Eds. 1988), and post-Variscan karstifications in North America and Europe, respectively (Głazek 1989a). Local (paleo)karst and products of Type 1 karst phases (*sensu* Głazek 1989a) are common products during the single unconformity karst 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.

The beginning and the end of karstification

Karstification of the host rocks *may start* during their formation phases – diagenesis – converting the soft sediment into consolidated material 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 into the diagenetic system. The formation of a fresh-water lens and a halocline zone related to the surface relief and sea-level changes is the result. The early stages of the origin of dissolutional (karst) porosity by meteoric diagenesis in carbonate rocks have been described in numerous sedimentological and paleokarst studies (a.o., Longman 1980; James & Choquette 1984; Tucker & Wright 1990; James & Choquette, Eds. 1988; Wrigh *et al*., Eds. 1991; Wright & Smart 1994; Moore 1989, 2001; Mylroie & Carew 2000). Some authors suppose karst to be merely the facies of meteoric diagenesis (Esteban & Klappa 1983).

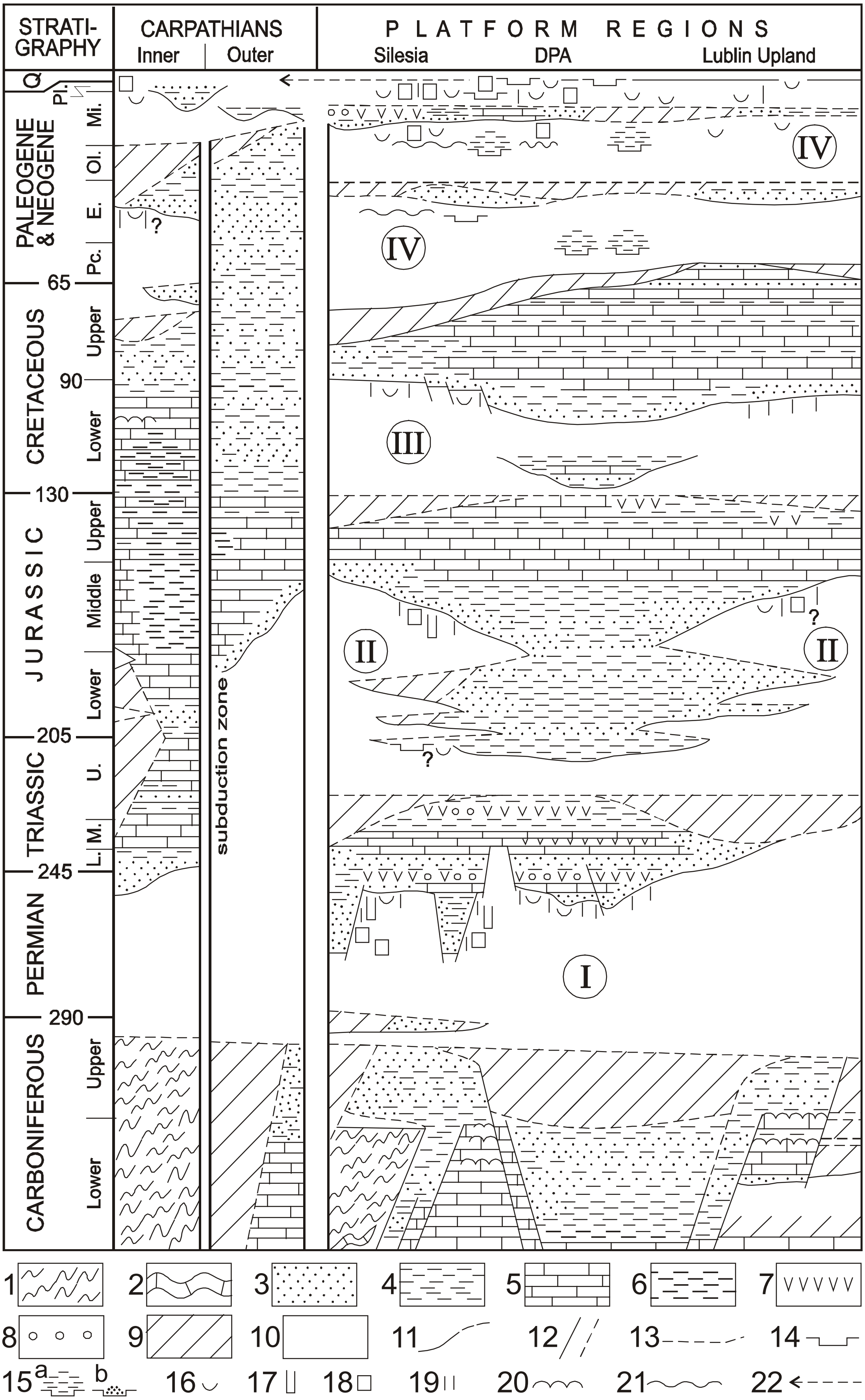
The *end* of karstification can be viewed also from various perspectives. An undisputed end of karstification occurs at the moment when the host rock, together with its karst phenomena, is completely eroded/denuded, i.e. at the end of the karst cycle *sensu* Grund (1914; see also Cvijić 1918). In such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

The evolution of karst is connected with chemical denudation, which defines the lowering of the surface of soluble rocks during a time unit. A review of this aspect is given by Ford & Williams (2007). The chemical denudation depends on the runoff. The calculated denudation rates resulting from dissolution vary from less than 0.01 to more than 760 $mm.a^{-1}$. Data on chemical denudation (surface lowering) have to be treated with caution. Data in Ford & Williams (2007) show that chemical denudation in the mild Central European climate vary between 15 and about 90 $mm.a^{-1}$, which corresponds to 15–90 m of dissolved limestone per one million of years. The latest data from paleomagnetic dating of cave infillings in the classical Karst of Slovenia show that the infilling of unroofed caves can be 6 Ma old (Bosak *et al.* 1998; Bosák *et al.* 2000) and maybe older. The oldest deposit dated until now occurs in the Račiška pečina Cave – speleothems about 3.2 Ma old (Bosák *et al.* 2004). If we accept values of chemical denudation for the region of the classical Karst in Slovenia (Gams 1981, 2003; see also Cucchi *et al.* 1994) of about 30–50 $mm.a^{-1}$, and we know that the caves originated at depths of at least 150–250 m (example of recent Škocjanske Caves), the speleogenetical phase of the presently unroofed caves (*sensu* Mihevc 1996) cut by the present surface should be 5 to maximally approx. 8 Ma old. The caves can, however, be even older, as the last preserved episode of infilling dates from 1.8 to about 6 Ma ago, and a cave deposit as old as 3.2 Ma in the Račiška pečina Cave is situated under a cave roof of tens of metres thick. Moreover, some data indicate that the present landscape could even have developed since the Early Badenian sea retreated about 15 Ma ago (Rögl 1998). It seems that the chemical denudation did not lower the surface in a regular way, or that lower values of 15–30 $mm.a^{-1}$ are more reliable. Similar discussions regard Australian karst (R.A.L. Osborne, pers. comm. 2006). In addition, the above values are valid for sedimentary limestones with a normal degree of lithification and diagenesis. Metamorphosed limestones (marbles) show quite different behaviour as they are less soluble, which results in a positive relief of limestone lenses in most of the crystalline terrains (common in the crystalline units of Moldanubia and in the Moravo-Silesian units of the Bohemian Massif, Czech Republic and elsewhere).

Time recorded in karst

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

The karst record of 1st–and 2nd-order stratigraphic discontinuities on the East European Platform and epi-Variscan Central European Platform in Poland was identified by Głazek *et al.* (1972) and Głazek (1973, 1989a). It encompasses a maximum of 50–60% of the geological time elapsed since deposition of the rocks (Fig. 1). Analysis of the Bohemian Massif (epi-Variscan Platform; Bosák 1987, 1997; Tab. 3, Fig. 2) showed that 12–45% of the geological time since the re-


STRATI-
GRAPHY
CARPATHIANS
Inner
Outer
P L A T F O R M R E G I O N S
Silesia
DPA
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Q
Pl.
Mi.
Ol.
E.
Pc.
PALEOGENE & NEOGENE
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Upper
CRETACEOUS
90
Lower
130
Upper
JURASSIC
Middle
Lower
205
U.
TRIASSIC
M.
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245
PERMIAN
290
Upper
CARBONIFEROUS
Lower
subduction zone
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Tab. 3. Review of temporal data for the evolution of the Bohemian Massif since the Paleozoic regression (after Bosák 1987)
Tab. 3. *Ewolucja Masywu Czeskiego po zakończeniu morskiej sedymentacji paleozoicznej (wg Bosáka 1987)*

Regional geological unit	Duration since Paleozoic 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

gression of Paleozoic seas in the Late Devonian/Early Carboniferous is represented in such records, and that 55–88% of the time is not recorded in the preserved marine or continental successions (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 resulting from the amalgamation of individual terrains 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 (Late Jurassic and Late Cretaceous regional transgressions, several minor Oligocene and Miocene transgressions reaching only the margins of the massif; see Fig. 2). The Polish territory is composed of slightly older elements in a different geotectonic setting, and the geological structure is little affected by younger orogenies. The platform cover was developed more continuously, and individual stratigraphic discontinuities represent shorter time intervals. Therefore, the preserved record of time differs significantly 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., the Scandinavian Shield), partly as a consequence of glacial isostasy. In such terrains, the time recorded can represent less than 10%. On some recent and fossil carbonate platforms, the time recorded in sediments represents only 5 to less than 10% (Quaternary Great Bahama Bank, Devonian carbonate platform on Moravia; Bosák *et al.* 2002).

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

Fig. 1. *Czas powstania form i osadów krasowych w Polsce (wg Głazka 1989b, zmodyfikowane)*
Zmetamorfizowane skały podłoża: 1 – skały krzemianowe, 2 – soczewy marmurów. Skały osadowe: 3 – piaskowce i zlepieńce, 4 – muły, iły, margle, 5 – skały węglanowe, 6 – siliciklastyczne i węglanowe skały głębokomorskie, 7 – siarczany, 8 – sole, 9 – skały zerodowane, 10 – degradacja subaeralna. Granice: 11 – niezgodności, 12 – uskoki synsedymentacyjne, 13 – przypuszczalny zasięg sedymentacji, 14 – polja, 15 – zagłębienia subrozyjne (wypełnione: a – węglem brunatnym, b – osadami klastycznymi), 16 – leje krasowe, 17 – studnie krasowe, 18 – jaskinie, 19 – małe formy krasowe, 20 – jaskinie syngenetyczne, 21 – powierzchnie erozji krasowej, 22 – maksymalny zasięg lodowców plejstoceńskich; I do IV – okresy aktywności krasowej, DPA – aulakogen duńsko-polski

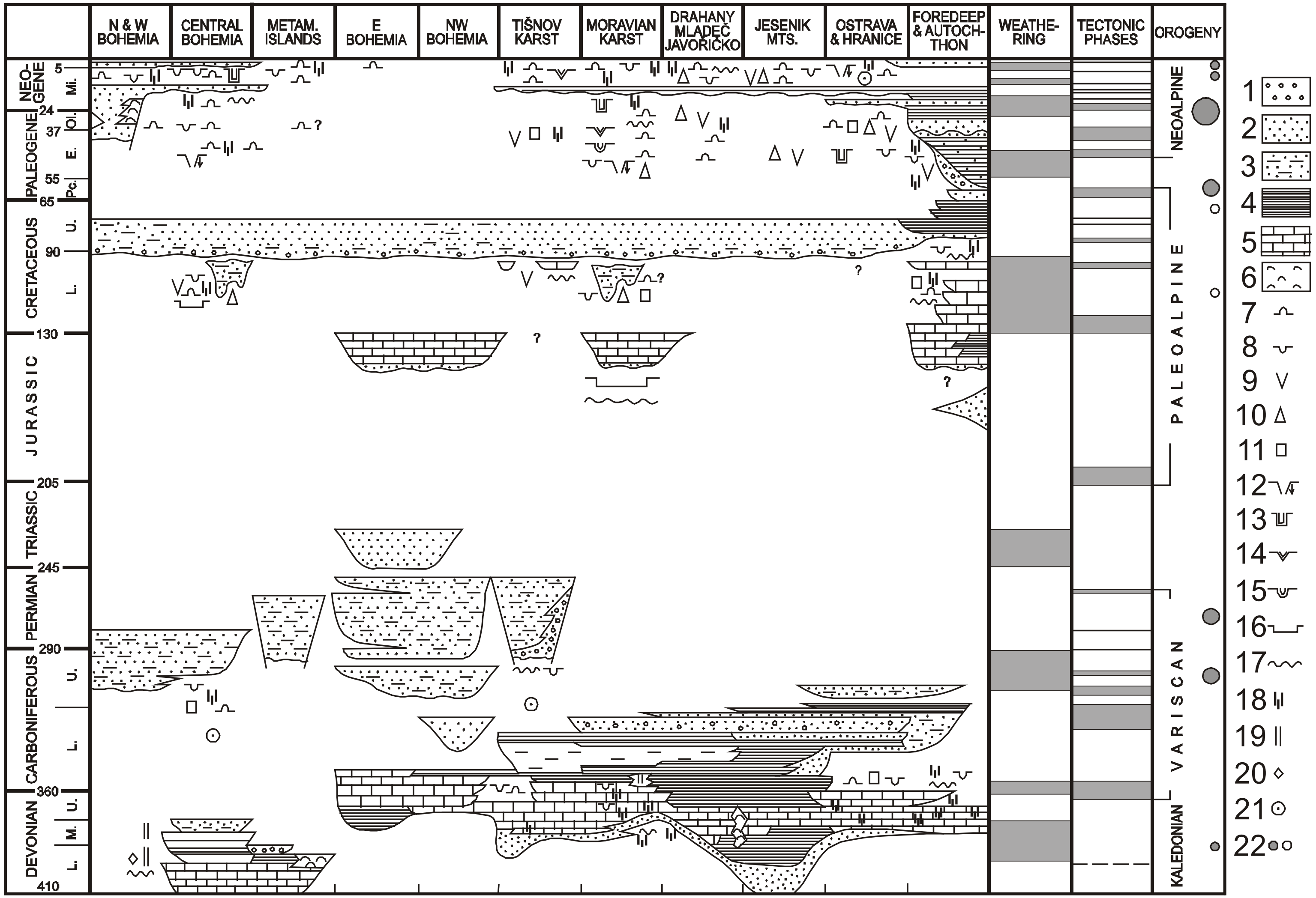

Fig. 2. Distribution of paleokarst and sediments in selected sections of the Bohemian Massif (simplified and modified after Bosák 1997)

Lithology: 1 – conglomerates, 2 – sandstones, 3 – lithologically variable siliciclastics (redbeds, sandstone-siltstone alternations), 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 diagenetic porosity, 21 – hydrothermal karst, 22 – volcanic activity: black – Bohemian Massif, white – Outer Western Carpathians adjacent to the Bohemian Massif, circle diameter approximately covers the time-span of volcanic activity

Fig. 2. *Rozprzestrzenienie form krasowych i skał osadowych w wybranych regionach Masywu Czeskiego (wg Bosáka 1997, zmodyfikowane)*

Litologia: 1 – zlepieńce, 2 – piaskowce, 3 – niejednorodne skały klastyczne (przemienne piaskowce i mułowce), 4 – łupki, 5 – skały węglanowe, 6 – skały wulkaniczne i wulkanoklastyczne. Formy krasowe: 7 – jaskinie, 8 – leje krasowe, 9 – żebra krasowe, 10 – mogoty, 11 – ostańce, 12 – studnie krasowe, 13 – wąwozy, 14 – doliny V-kształtne, 15 – doliny U-kształtne, 16 – polja i duże obniżenia krasowe, 17 – powierzchnie erozji krasowej, 18 – małe formy krasowe, 19 – dajki neptuniczne, 20 – porowatość w wyniku działalności wód meteorycznych, 21 – kras hydrotermalny, 22 – okresy aktywności wulkanicznej: czarne kółka – w Masywie Czeskim, białe kółka – w Zewnętrznych Karpatach Zachodnich, średnica kółka odpowiada długości trwania aktywności wulkanicznej

Products

It can be readily deduced 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 geomorphological processes. The longer is the duration of subaerial exposure, the more complex are these geomorphological agents. Some processes can destroy karst features in a relatively short time, leaving planated surfaces with little or no traces of previous karstification, e.g. the effect of marine transgressions (represented by an unconformity of the 3rd order). This can be illustrated for recent karst in the coastal zone of Palawan Island (Philippines) and the Lower Devonian of the Koněprusy area, Czech Republic. On Palawan, Longman & 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 rare relics of truncated dissolution fissures and sinkholes. An identical situation is detected at the boundary between the Koněprusy Limestones (Pragian) and the Suchomasty Limestones (Dalejan, Early Devonian) at Koněprusy. The truncation plane, which is well exposed in the 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.

Products of longer subaerial exposure (unconformity of the 3rd to 2nd order) of carbonate rocks can be illustrated for the Early to Middle Cretaceous polycyclic evolution of an island system in the Tethyan realm. Mišík (1978), Mišík & Sýkora (1981) and Aubrecht *et al.* (2006) reconstructed the emerged relief of the Pieniny cordillera (Slovakia) with a width of several tens of kilometres and a diversified relief, including river basins and karst-affected limestones. The karst was destroyed during the Late Cretaceous, and relics of speleothem and fresh-water limestones were deposited in post-Santonian conglomerates. Recent equivalents can be seen in (1) the Greek archipelago, which is composed mostly of parts of emerged limestone platforms; (2) the Indonesian-Philippines island arc, and (3) Papua-New Guinea with highly evolved karst landscapes and cave systems. But there, too, the evolution of cave systems and levels took a substantial part of the geological time (a.o., Noel & Bull 1982; Audra *et al.* 1999).

Products of paleokarst evolution are best preserved directly beneath a cover of marine or continental sediments, i.e. under the deposits that 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. Therefore, the ages of particular paleokarsts have most commonly been assigned to times shortly before the termination of the stratigraphic gap (Bosák 1997). This fact can be easily illustrated for pre-Cenomanian paleokarst in the Bohemian Massif, for pre-Callovian karst in Moravia and for Westphalian/Stephanian karst in central Bohemia (see Fig. 2). An identical situation occurs in Poland (a.o., Głazek 1989b; Paszkowski 2000; see Fig. 1)

Tab. 1 summarizes some karst forms, products and processes related to unconformity order against the background of transgression/regression sets during a hypothetical karst period (which can be compared with Figs. 1 and 2). The conformities, with no measurable time gap but with a distinct short interruption in deposition, can be characterized only by diagenetic changes (see a.o., Longman 1980; Tucker & Wright 1990; Moore 1989, 2001), burrows of organisms (trace fossils – a.o., Bromley 1975) or the formation of hardgrounds (a.o., Bathurst 1971; Wilson & Palmer 1992).

As stated above, the more time is available, the more evolved karst landscapes can form. Moreover, the higher is the difference in elevation between the base level and the highest summits, the deeper and more evolved a vadose zone can be formed. This implies that slightly and shortly emerged carbonate sediments above sea level can show slight karstification (meteoric diagenesis) in the form of different *karren* types: like in the Carboniferous of Great Britain (a.o., Ramsbottom 1973; Walkden 1974, 1977; Wright 1988). *Soils* are also connected with low-order unconformities; examples are the protosols in the Quaternary of the Bahamas formed during sea-level falls in oxygen isotopic stage 5 (a.o., Panuschka *et al.* 1997), and the more complex soils in the Carboniferous of Great Britain (a.o., Adams 1980; Bridges 1982) or Devonian of Holy Cross Mountains (Poland, Skompski & Szulczewski 2000). *Caliche, carbonate duricrusts and hardened zones* can be formed in a semi-arid (or distinctly seasonal) climate under conditions with a vadose zone (*cf.* Esteban & Klappa 1983) in relatively short time-spans, for instance like during a single sea-level fall within the oxygene isotope stage 5 on San Salvador Island, Bahamas (Bosák, Hladil & Slavík unpubl.), or in the Carboniferous of Great Britain (a.o., Wright 1982). Evolved *weathering crusts* are connected with high-order unconformities and disconformities; it is estimated that one metre of laterite or bauxite needs about 0.4–1 Ma to develop (*cf.* Bárdossy 1982; McFarlane 1983). Nevertheless, the transported products of long-lasting weathering can be deposited on a fresh karst relief on emerged carbonate platforms attached to cratons or platforms (a.o., Bárdossy 1982, 1989; Bourrouilh-Le Jan 1989). The thickness of a karst infilling depends on the position of the groundwater table, which is related to the altitude of the emerged strata (a.o., Bárdossy 1989).

Each emergence forms conditions for the origin of a *fresh-water lens* and a fresh/salt-water mixing zone (halocline; see Mylroie & Carew 2000 for an excellent review) and for the evolution of a complex set of fresh-water, mixed and marine diagenetic processes in the vadose and phreatic zones (*cf.* Longman 1980; Tucker & Wright 1990; Moore 1989, 2001) and of special karst forms related to this complex environment (Mylroie & Carew 2000; Gunn & Lowe 2000).

Although *hydrothermal activity* is usually connected with the general karst model (Esteban 1991), it can also be detected in some fossil examples of the Caribbean model. Hydrothermal activity connected with the expulsion of basin waters affected Early Devonian (Lochkovian and Pragian) limestones of the Koněprusy area (Barrandian, Czech Republic). While the Lochkovian strata are dissected by a dense network of calcite veinlets, the Pragian formations are impregnated with organic matter (bitumen: Franců *et al.* 2001), phosphate and silica resulting from the transport of organic matter, from dephosphoritization of conodonts and from desilicification of sponge spicules in the underlying Lochkovian and Late Silurian sediments. Later, the hydrothermal springs became enriched in Mg and S, and metasomatic spots with crystallization of dolomite, illite and pyrite were formed (Hladil & Gabašová 1993). The hydrothermal activity was connected with hot springs (100–130°C) ascending along the transpression fault zone and lasting about 2.5 Ma, which is the duration of the hiatus between the Lochkovian and Pragian strata in some tectonic blocks in this part of the Prague synform (Hladil 1997; Hladil & Slavík 1997).

Time for evolution of a conduit

The evolution of a conduit is a rather complicated set of events facing numerous critical thresholds (for a summary, see White 1988; Palmer 2000, 2002; Ford & Williams 2007). Three phases of speleogenesis are now generally accepted: (1) initiation: initial enlargement of a fracture to a critical size; (2) breakthrough: a fairly sudden transition to rapid dissolution, resulting in the growth of an incipient cave into a true cave, and (3) enlargement: the growth of a protoconduit/incipient cave to full conduit size (a.o., White 1988; Palmer 2002).

The initial fracture permeability and/or rock porosity has connected apertures of the order of 50–500 μm, and the diameter of a solution protoconduit reaches 5–15 mm (White 1988, Ford & Williams 2007). When the diameters reach a size of 0.5–5 cm, a kinetic breakthrough occurs (Dreybrodt & Gabrovšek 2000) and the flow may change from laminar to turbulent (White 1988; Ford & Williams 2007), enabling the transport of detrital sediment (Palmer 2002).

Initiation phase

The duration of a typical initiation phase was calculated to be approx. 3–5 ka (White 1988), based on experiments by Howard & Howard (1967) and calculations of Palmer (1981). They stated that the maximum dissolution rate is 0.14 $m \cdot a^{-1}$. Palmer (1991) calculated the initiation phase to last minimally 10 ka under favourable conditions. Dreybrodt & Gabrovšek (2000) estimated the duration of the initiation (gestation) phase for realistic cases to be 1 ka to 10 Ma. The time depends critically on the length and the initial width of the fracture. Palmer (2002) assumed that the enlargement of initial openings to cave size would require many millions of years, except under the most ideal conditions.

Breakthrough phase

The duration of the breakthrough (gestation time) has been discussed by Palmer (2002). The time necessary for the process is influenced by numerous parameters (like hydraulic gradient, temperature, P_{CO2}, organic and other acids) and amounts to 10^4 to 10^5 years, although Dreybrodt (1990, 1996), Dreybrodt & Gabrovšek (2000), Bauer *et al.* (2005) estimate the time required to be shorter.

Enlargement phase

The enlargement phase, i.e. the time in which the protoconduit develops to full size, is affected by numerous thresholds (see a.o., White 1988) and agents, including geological conditions (lithology, primary and secondary porosity), climatic conditions (temperature, precipitation, water volumes), hydrochemical conditions (concentration and kind of solvent agents), and clastic load in the cave waters (its transport and deposition can enhance or retard dissolution and erosion: Palmer 2002). All these conditions affect the velocity of speleogenesis. The estimates provided underneath are therefore to be considered only as approximations.

The time of transition from protoconduit to traversable cave (with a diameter of 1–10 m or more) is expected to be 5–20 ka up to 100 ka in many geological settings (White 1988). Ford & Williams (2007) 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, 2000, 2002) calculated the maximum wall retreat to be 0.001–0.15 $cm \cdot a^{-1}$ in a typical meteoric groundwater cave; water-filled caves thus might increase their diameter from 0.2 to 2–3 m in 1 ka, depending on the hydrochemical conditions. Dreybrodt & Gabrovšek (2000) estimated the velocity of enlargement of a conduit under phreatic conditions to about 200 $mm \cdot ka^{-1}$, so that a phreatic passage of 30 m in diameter can be developed within 100 ka. For hydrothermal caves, durations of the order of 10^5 to 10^6 years are required to produce caves of traversable size (Palmer 1991). The development of each passage level in Mammoth Cave (Kentucky, USA) required at least 10^5 years (Granger *et al*. 2001), which includes time for breakthrough and for later enlargement to the present diameters of about 5–10 m in the major passages (Palmer 2002). Mylroie (1977) described the formation of traversable passages up to a metre in diameter and 200 m long since the last deglaciation at about 13 ka ago. Data of Ford (1980) and Palmer (1984) suggest that an extension time of 10–100 ka per kilometre of the conduit may have prevailed in a majority of karst settings. White (1988) obtained an extension rate of 3–5 ka per kilometre. Vadose entrenchment in canyons in caves of New York State (USA) was measured to be 10–20 $mm.ka^{-1}$ (Palmer 1996).

Theoretical assumptions have been proven by field observations. Mylroie & 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), so that 36 ka was available for the cave formation along the halocline. Numerous data from North America and Ireland indicate a post-glacial origin of caves that are perfectly adjusted to recently deranged surface landscapes and hydrologic regimes, i.e. the caves developed during the last 8–15 ka (a.o., Mylroie 1977; Mylroie & Carew 1986, 1987; White 1988; Ford & Williams 2007).

Non-carbonate karst

The above-mentioned discussion focused on karst in carbonate rocks (limestones), which do, however, not represent the most soluble rock types in the Earth crust. Sulphates (gypsum) and halites (rock salt) represent also substrates on which karst is extensively developed. Nevertheless, true karst is developed also in poorly soluble rock types, like quartzose sandstones and quartzites (for an excellent review, see Wray 1997).

Karst in evaporites

The solution of evaporites is due to a simple interaction of rock and water. In contrast to carbonate rocks, no other substance enhancing the solution rate is necessary to be added: the system is composed of two agents (so-called parakarst of Cigna 1986). The dissolution rates of gypsum and anhydrite are much (about 2 orders of magnitude) higher than those of calcite and dolomite (for reviews, see a.o., Ford & Williams 2007; Klimchouk *et al.,* Eds. 1996; Klimchouk 2000). Halite (rock salt) is much more soluble than sulphates (about 140 times) (Klimchouk *et al.*, Eds. 1996; Klimchouk 2000; Frumkin 2000).

The chemical denudation on sulphate surfaces is roughly 10 times higher than on carbonates (Ford & Williams 2007). The regional karst denudation rate on rock salt in the arid to semiarid climate of Israel (Mount Sedom) is about 50–75 $m.ka^{-1}$ (Frumkin 1994, 2000), which is 1–2 orders of magnitude higher than limestone denudation rates in more humid areas. Evaporite minerals tend to be dissolved in the deep underground (*subrosion*) in huge quantities, forming special types of landscapes (e.g., zero subedge or salzhange: Ford 1989; subrosion „maarlike“ depressions, often coal-bearing: Meiburg 1980).

It is evident that the time necessary for the development of a karst landscape on sulphates and rock salt is shorter than on carbonates; it is taken top be several thousands or tens of thousands of years. Caves can form within the same time period, or even more rapidly. Pošepný (1893, 1902) described sudden floods in deep salt mines in Romanian salt plugs. The floods entered the mine by cave conduits in the salt; these conduits had developed within some tens of years, connecting a surfacial stream with the artificial underground caverns. Observations during some ten years in the longest salt cave of the World (the 3N Cave in the Namakdan salt plug, Qeshm Island, Iran) indicates that each flooding is responsible for a change of the cave morphology (Bruthans *et al.* 2006a; Filippi *et al*. 2006). The cave developed after a regression of the sea about 5 ka ago (Bruthans *et al.*, 2006b). The downcutting in the caves of Mount Sedom (Israel) takes 8 orders of magnitude faster place than in limestone caves (Frumkin & Ford 1995).

The high velocity of the dissolution processes in evaporite areas allows the karst to evolve during relatively short time-spans, related to low-order unconformities.

Karst in siliceous rocks

Karst in siliceous rocks also represents parakarst (*sensu* Cigna 1986). The process is based on slow dissolution (hydration) of opal and/or quartz cement of clastic grains (Martini 1979) or on direct quartz dissolution along fissures, cracks and grain contacts (Wray 1997), followed by a suffosion and piping (erosional removal of dissintegrated rock). Nevertheless, owing to the extremely low solubility of quartz under natural conditions (5–25°C; Siever 1962), the evolution of karst in sandstones and quartzites demands substantial time (many millions to hundreds of millions of years) and conditions of intensive weathering under tropical settings (Martini 2000); this implies a karst type related to stacked high-order unconformities.

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