Chapter I.5

SPELEOGENESIS IN GYPSUM

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Satisfactory explanation of the origin and development of caves (speleogenesis) is a core problem of karst studies. Karst evolves as a circulation system, organised and interconnected through a conduit structure. Such a system may include superficial inputs and outputs, expressed as or related to karst landforms. However, there may be no such components if the system is represented entirely by conduits as in the case with deep-seated intrastratal karst.

The main differences between speleogenesis in gypsum and in carbonate rocks lie in the chemistry and kinetics of their dissolution, in some of the lithological or structural peculiarities of the respective rocks and formations, and in their hydrogeological characteristics. These sets of factors are examined in detail in chapters I.1, I.2, and I.6 respectively. The present chapter considers how these factors influence cave origin and development.

1. Caves in gypsum karst

Currently there are perhaps several thousand gypsum caves known around the word. In terms of the karst typology adopted here (see Chapter I.4), most of the caves that can be explored directly are found in exposed and intrastratal entrenched karsts. Caves are found more rarely in intrastratal subjacent karsts and are almost never found in deep-seated intrastratal karsts. The latter two karst types are by far predominant in terms of areal extent. It can be assumed that known caves represent only a very small portion of all the karst conduits and voids that occur within the upper few hundred metres of the geological sequence in gypsum karst areas. However, to justify the above statement, it must be demonstrated that caves are as common in deep-seated intrastratal karsts as they are in the entrenched and exposed types.

The world's largest currently known gypsum caves are listed in Table 1 Optimisticheskaja, the longest cave, is the second longest cave of any type known in the world. A striking gap exists between lengths of the three longest caves and the other caves in the list. The five longest gypsum caves, located in the Western Ukraine, account for well over half of the total known length of gypsum caves. This apparent bias is related partly to the unique structural prerequisites of speleogenesis, which are locally realised under artesian conditions. It also reflects a favourable regional evolution (with rapid uplift, and fossilization of labyrinth systems), the presence of overlying limestones, and considerable clayey protective cover (which prevented the infilling and/or destruction of the huge mazes). It is far more common, however, that artesian (intrastratal) caves in gypsum are partially destroyed while passing from conditions of intrastratal karst to those of entrenched and exposed karst. Moreover, all genera of caves in gypsum , whether relict or newly-formed, are more readily destroyed in exposed and shallow sub-surface environments than are those in carbonate karsts, due to the lower mechanical strength and greater inhomogeneity of gypsum forma-

Table 1a

Longest gypsum caves of the world (as for 1996)			
Name	Development, m	Country	Rock age
Optimisticheskaja	200000+	Ukraine	Neogene
Ozernaja	117000	Ukraine	Neogene
Zolushka	92000	Ukraine	Neogene
Mlynki	25000	Ukraine	Neogene
Kristalnaya	22000	Ukraine	Neogene
Kulogorskaja-1-2 - Troja	14100	Russia	Permian
Jester	11800	USA	Permian
Spipola-Aquafredda	10400	Italy	Neogene
Slavka	9100	Ukraine	Neogene
Agua, cueva de	8350	Spain	Neogene
Verteba	7820	Ukraine	Neogene
Cater Magara	7300	Syria	Neogene
Park's Ranch	6269	USA	Permian
Konstitutzionnaja	5880	Russia	Permian
Kungurskaya Ledjanaya	5600	Russia	Permian
Olimpijskaja	5500	Russia	Permian
Kumichevskaja	5000	Russia	Permian
Zolotoj Kljutchik	4380	Russia	Permian
Covadura	4245	Spain	Neogene
Crystal Caverns	3807	USA	Permian
Double Barrel	3724	USA	Permian
Scrooge	3700	USA	Permian
Umm al Masabih	3593	Libya	Jurassic
Leningradskaja	3400	Russia	Permian
Simfonia	3240	Russia	Permian
Pedro Fernandes	3204	Spain	Neogene
Pekhorovskaja	3180	Russia	Permian
Martin	3150	USA	Permian
Lomonosovskaja	3127	Russia	Permian
Carcass	2920	USA	Permian
Vodnaja	2900	Russia	Permian
Pinezhskaja Terehchenko	2600	Russia	Permian
Atlantida	2525	Ukraine	Neogene
Ingh. Ca' Siepe	2500	Italy	Neogene
Wimmelburger Schlotte	2840	Germany	Permian
Eras'kina 1-2	2500	Russia	Permian
10-years LSS	2450	Russia	Permian
Bukovinka	2408	Ukraine	Neogene
Hyaenenlabyrint	2310	Somaly	Paleogene
Severjanka	2300	Russia	Permian
Pekhorovskij Proval	2266	Russia	Permian
Kulogorskaja-5	2200	Russia	Permian
Geograficheskogo ob-va	2150	Russia	Permian
Ugryn'	2120	Ukraine	Neogene
Re Tiberio	2110	Italy	Neogene
Gostry Govdy	2000	Ukraine	Neogene

Table 1b: the world's deepest (>100m) gypsum caves (as of 1996)

The deepest gypsum caves of the word			
Name	Denivelation, m	Country	Rock age
Tunel dels Sumidors	210	Spain	Triassic
Pozzo A	>200	Italy	Triassic?
Corall, sima del	130	Spain	Neogene
Triple Engle Pit	130	USA	Permian
Covadura	126	Spain	Neogene
Campamento, sima del	122	Spain	Neogene
Aguila, sima del	112	Spain	Triassic
Rio Stella-Rio Basino	100	Italy	Neogene
AB 6	100	Russia	Jurassic

Note: The list has been compiled using data provided by Belski (for USA), Calaforra (for Spain), Forti & Sauro (for Italy), Kempe (for Germany), Klimchouk (for Ukraine), Woigt & Schnadwinkel (for Syria), Malkov & Lavrov (for Russia), Ehrsam (for Somaly)..

tions. These reasons also account for the relative scarcity of 20-80km-long gypsum caves, when compared with the full class of limestone caves. They also explain the generally much more modest sizes of the biggest chambers and passages in gypsum compared to those in limestones.

The common occurrence of caves under currently deep-seated artesian conditions is proven in many regions of intrastratal karst. Boreholes and mines have intersected large voids at depths below local base levels of 60 -100m (in the "artesian belt" of the gypsum karst in the Western Ukraine; see Chapter II.9), 300 - 400m (in the Pre-Urals, the Caspian depression, Russia; and the South Hartz region of Germany), or even deeper. Such deep-seated development is also evidenced by the presence of collapse forms, which have evolved after vertical through structures (see Chapter I.9) where gypsum lies at depths of several hundreds meters.

The deepest gypsum cave currently known (Tunel dels Sumidors, Valencia, Spain; see Chapter II.6) is only 210m deep, far shallower than the deepest caves in carbonate karsts. The main reasons are geological. In folded, mountainous regions, where the potential drained depth is greatest, gypsum formations are fragmented and do not favour the development of such vertically extensive sequences as do carbonates. Again, the lesser mechanical resistance and homogeneity of gypsum formations restricts the possibilities of deep gypsum caves developing and surviving.

Gypsum caves vary greatly in morphology. Particularly on the level of system patterns this commonly reflects, genetic differences. Several typical patterns can be distinguished:

1) Discrete, comparatively large voids, often isometric.

2) Rectilinear or ramifying mazes. Multiple storeys may complicate the structure;

3) Caves that are linear or crudely dendritic in plan and horizontal, inclined or step-like (with pits) in profile. Multiple storeys may complicate the structure;

4) Vertical pipes.

Development of these types of caves is related to particular speleogenetic environments and karst types (Table 2). Complicated evolution of karst systems may cause superimposition of diffe-

Table 2.

Genetic classification of caves in gypsum

TYPE OF KARST	SPELEOGENETIC SETTINGS			CHARACTERISTICS OF CAVES
	Hydro-geological conditions principal complementary	Flow pattern through gypsum and type of recharge	Initial permeability (before speleo-genesis	
Intrastratal deep-seated	confined (artesian)	1. Ascending flow, localized basal inputs	very inhomo-geneous, generally low to negligible, locally high	1. Discrete voids, com-monly large and isometric; associated stoping cavities in coverbeds at the top of VTS after breakdowns
		2. Ascending flow with possible lateral component, dispersed basal inputs	fairly homogeneous, generally low	2. Rectilinear 2-D or 3-D (multi-storey) mazes
Intrastratal subjacent	confined phreatic water table vadose	1. Ascending flow with possible considerable lateral component, localized or dispersed basal inputs	heterogeneous: low to high	Continuing development of the types 1 and 2
		2. Descending flow with considerable lateral compo-nent, localized inputs from coverbeds and via super-ficial sink points; possible backflooding		3. "Through caves": caves that are linear or crudely dendritic in plan, horizontal, inclined, or step-like in profile Enlargement of inherited caves at the water table
Intrastratal entrenched	phreatic water table vadose	1. Descending flow with possible considerable lateral component, localized inputs from coverbeds and via superficial sink points; possible backflooding	heterogeneous: low to high	Continuing or newly started development of the type 3 caves 4. Vertical pipes developing downwards from the top of the gypsum
Exposed denuded	phreatic water table vadose	1. Descending flow with possible considerable lateral component, localized inputs from coverbeds and via superficial sink points; possible backflooding		Continuing or newly started development of the type 3 caves
Exposed barren	phreatic water table vadose	1. Descending flow with possible considerable lateral component, localized inputs from coverbeds and via superficial sink points; possible backflooding		Continuing or newly started development of the type 3 caves

rent features and structures. Detailed characterisation of the above types of caves and their corresponding speleogenetic environments is given in sub-chapter 6 below. The suggested draft classification does not encompass all caves occurring in gypsum, but covers only those created by underground water circulation imposed upon aquifers. Other types include, for instance, cavities formed due to differential deformation of layers due to recrystallization ("tumulus"), or gravitational/tectonic caves formed, for example, due to unloading along escarpments.

2. Hydrographic (hydrodynamic) zones, artesian versus phreatic conditions

A long lasting controversy concerning the validity of vadose, water table and phreatic theories of cave development has been resolved during recent decades. The "four state" model of Ford & Ewers (1978) clarified the role of each corresponding environment. It is well accepted that most conduits originate under phreatic conditions, although their development or modification may continue at the water table or within the vadose zone (Palmer, 1984; Ford, 1988; Lowe, 1992). Some speleoforms may develop entirely under vadose conditions. However, a recently elaborated theory (Klimchouk, 1990, 1992, 1994, 1997a) suggests that caves may also originate and develop under artesian (confined) conditions and be subsequently modified in the phreatic, water table and vadose environments. Acceptance of these ideas allows the common confusion concerning the meanings of the terms "phreatic" and "confined" to be clarified.

In contrast to water-table or unconfined aquifer conditions, where the water table is under atmospheric pressure, the water pressure in a confined aquifer is greater than atmospheric at any point, as the head in such an aquifer is above the bottom of the upper low permeability confining bed. Any breaching of the upper confinement, such as by a well, a fault or a facial "window", will cause water to flow upwards to the level where the water column is high enough to balance the aquifer pressure (potentiometric level). This effect is mainly caused by water entering an aquifer at elevations greater than that of the base of the confining bed across most of the aquifer's areal extent, though there are other possible sources of pressure generation. Several confined aquifers may exist in a system, separated by poorly permeable beds. Confined aquifers are commonly called artesian aquifers, and confined conditions commonly referred to as artesian conditions.

The term phreatic implies conditions where water saturates all voids in a rock, in contrast to vadose conditions, above the water table, where voids are only water-filled transiently. Water in phreatic conduits is always confined by the host rock and possesses some hydraulic head above the conduit ceiling. This has given rise to some confusion where the terms "phreatic" and "artesian" ("confined") have been wrongly understood as being equivalents, especially when considering deep phreatic conditions. For example, Glennie (1954) termed water rising from such deep phreatic paths "artesian". Jennings (1971, p.97) noted that such usage is in a strict sense incorrect, but it serves as a reminder that consolidated rock can act virtually as its own aquiclude. It is necessary to distinguish the term "artesian" ("confined") as referring to flow conditions in a whole aquifer (or a system of aquifers where there is major geological confinement), rather then to flow conditions within a single conduit. Use of the term "phreatic" should be restricted to description of the lower zone in an unconfined aquifer, limited above by a water table that is free to rise and fall.

In hydrogeological terms, flow in artesian aquifers is considered to be: "...in many ways an extreme example of the effects found in the phreatic zone of unconfined aquifers, with the vertical hydraulic gradients increased as a result of the presence of the overlying confining bed." (Price, 1985, p.68). However, the fact that the distinction between phreatic and confined conditions is of ultimate importance for speleogenesis was not fully recognised in the past. The main difference is that in phreatic conditions discharge through a potentially developing flow path is governed by the resistance of the path itself, particularly of its narrowest part. In confined conditions discharge through conduits is governed by the resistance of the least permeable bed that causes major confinement of a system in the vertical down-gradient direction. This point is examined more fully in sub-chapter 5 below.

3. Implications of gypsum dissolution chemistry and kinetics

The chemistry and kinetics of gypsum dissolution have been considered generally in Chapter I.2. Important peculiarities affecting speleogenesis are:

1. The solubility of gypsum in pure water (2.531 g/L at 20°C) is roughly 10-20 times greater than the solubility of calcite in the presence of CO_2. In most relatively shallow environments (intrastratal entrenched karst, exposed karst) the influence of temperature variations is minor, but the effects of ion pairing, which increases gypsum solubility by up to 10%, must be considered.

2. Most commonly in deep-seated environments, but in subjacent karst settings too, several chemical and physical factors may (and in many regions are recorded to) increase or renew gypsum solubility considerably. The most important of these are: the presence of other salts in groundwaters (which enhances ionic strength and increases gypsum solubility by up to 3 times); anaerobic reduction of sulphates in the presence of organic matter; de-dolomitization of intercalated dolomite layers; and stress applied to the rock.

3. Whereas the kinetics of gypsum dissolution are described by the first order equation, anhydrite dissolution rates obey the second order equation. For gypsum, the flow time (distance) at which dissolution approaches 90% of saturation is very short; the dissolution rate decreases several orders of magnitude above this limit, and the 100% saturation level is approached asymptotically

4. Gypsum and anhydrite dissolution proceed even in contact with static water, but dissolution rates increase rapidly with increasing flow velocities. As a turbulent flow regime sets in, dissolution rates are boosted, probably by an order of magnitude.

5. The presence of other salts, such as sodium chloride, in solution considerably increases gypsum dissolution rates, but influences anhydrite dissolution rates even more drastically. This effect contributes particularly to speleogenesis in deep-seated environments.

4. Structural and hydrostratigraphical pre-requisites of speleogenesis

Initial permeabilities of common aquifers (e.g. some clastic rocks) in deep-seated settings is normally greater then that of karstifiable units, particularly sulphates, before the onset of speleogenesis. Gypsum beds commonly act initially as separating beds. Groundwater comes into contact with gypsum either from adjacent aquifer formations, or via minor beds of other lithologies, such as marls or dolomites, that are intercalated with the gypsum beds. This view is somewhat similar to that suggested by Lowe (1992) in his "Inception Horizon Hypothesis", although the intercalations in gypsum sequences do not commonly generate specific dissolution chemistry, as is suggested for carbonate sequences. However, such horizons in gypsum formations may locally determine chemical processes that maintain the gypsum dissolution potential, due to removal of sulphates from solution (sulphate reduction, de-dolomitization).

Late diagenesis (catagenesis) and/or tectonism impose fissure permeability upon gypsum formations, which then begin to play the major role in determining initial flow paths through the gypsiferous sequence. Gradient fields in confined conditions generally promote vertical, cross-formational circulation, although the presence of water-conducting intercalations and a specific fissure distribution may support a significant lateral component of speleogenetic development. While major tectonic fissures normally cut through the entire thickness of a bed, lithogenetic fissures tend to form largely independent networks that are confined within certain textural intervals. Such networks are characterised by a good lateral connectivity, but are connected vertically only at a relatively small number of discrete points. This provides the structural pre-requisites for the development of multi-storey maze caves (Klimchouk, 1992, 1994; Klimchouk et al., 1995; see also Chapter I.1).

5. Origin and development of conduits in confined and phreatic conditions

5.1. Origin and propagation of early conduits

A theoretical approach to the understanding of the propagation of early dissolutional openings in fissures has been developed by Palmer (1984, 1991), based on the combined consideration of mass-balance relationships, hydraulic equations for laminar flow, and chemical mass-transfer. Dissolutional enlargement of partings in soluble materials, gypsum in particular, has been investigated theoretically and experimentally by James & Lupton (1978) and James (1992).

It is generally believed that most proto-caves propagate through fissures where the connected apertures are small: limits between $<10\mu m$ to 1mm have been suggested (Ford & Williams, 1989). Recent study by Groves & Howard (1994) suggests that a minimum aperture of $100\mu m$ is required for conduit development. Seepage through them is very slow and laminar. The rate and configuration of dissolutional widening depends primarily on discharge through the fissure and the change in solute concentration along its length (Palmer, 1984, 1991; James, 1992). Due to gypsum's fast dissolution kinetics, solute concentration increases rapidly to about the 90% saturation level, so that the penetration distance, L_{90}, is quite short (for details of the L_{90} concept see Weyl, 1958; White, 1977; Ford & Williams, 1989). Fissures enlarge at their inlets, remaining almost unchanged downstream resulting in a tapered geometry. The mode of dissolutional enlargement of the fissure, or through a sequence of inter-connected fissures, will change only when a breakthrough of the penetration distance to the output boundary occurs, so that a flow path enlarges enough to permit an increase in flow velocity and penetration of significantly undersaturated water beyond the exit.

It is not yet clear whether the main breakthrough mechanism is the propagation of a taper

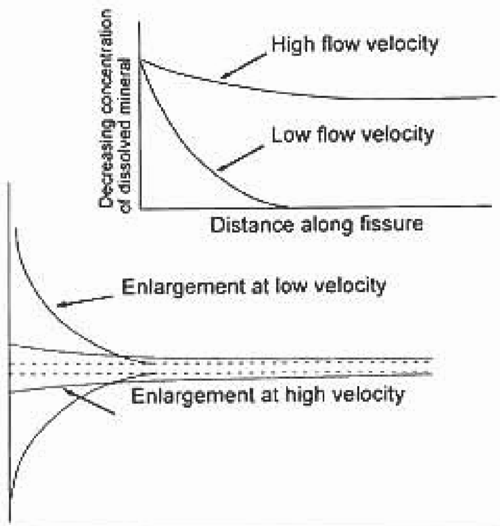

Fig. 1. Effect of flow velocity on enlarge-ment configuration (After James & Lupton, 1978).

from the input end, or slow dissolutional enlargement along the entire fissure length by water close to saturation. Shapes of enlarged fissures depend upon rock solubility, the dissolution rate constant, and flow velocity. Other variables being equal, fissure enlargement at the input end will be more tapered in gypsum than in limestone due to faster dissolution kinetics. Lower initial velocities produce sharp tapers, whereas higher velocities promote a more gradual enlargement (Fig.1). The sensitivity of the enlargement configuration to flow velocities, and hence to differences in the initial fissure width, should be more pronounced in fissures in gypsum than those in limestone.

Under sluggish flow conditions in gypsum it seems possible that the tapering is so localised that the first part of a fissure may reach human-penetrable sizes, while its downstream parts remain almost unchanged and very narrow. The common occurrence of "blind" ends of dissolutional passages in the labyrinthine caves of the Western Ukraine, with narrow guiding fissures recognizable along their apex, supports such an assumption. This introduces some potential confusion of the morphometric criteria that allow distinction between the inception and development stages in gypsum speleogenesis.

The duration of the initiation phase (until breakthrough is achieved) depends mainly on the length of fissure paths and on initial flow rates; the latter in turn depends strongly upon the width of the initial aperture (it is proportional to the cube of the width). This dependence is the primary factor responsible for the selective enlargement of openings (Fig. 2). The effect is more pronounced in gypsum than in limestone due to diffusion control of dissolution kinetics, which causes stronger dependence of the dissolution rate constant upon flow velocity. It is quite possible that duration of the initiation phase of conduits propagating through narrow fissures in gypsum is comparable with, if not longer, than that in carbonates, given that the boundary hydraulic conditions are equal. This apparent paradox is caused by dissolution kinetics being faster in gypsum than in carbonates.

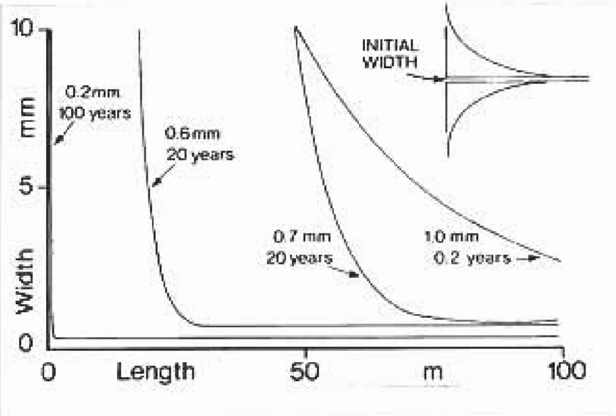

Fig. 2. Penetration distances or progress of the dissolution front for - L_{90} in massive gypsum, calculated for initial fissure widths ranging from 0.21 - 1.0mm. Time elapsed since initiation is in years. The hydraulic gradient is 0.2 and water temperature is 10°C. Inset: the form of the dissolution taper into the fissure that is obtained from theoretical calculations such as these (After James & Lupton, 1978, as adapted by Ford & Williams, 1989).

5.2. Development of conduits

At this point the distinction between phreatic and confined conditions becomes crucial to speleogenetic development.

Phreatic conditions. The controls of conduit development under phreatic conditions are best described by Palmer (1984, 1991). Given a substantial hydraulic gradient, the amount of flow through a fissure path is determined by its width. During the initiation stage water emerging at the output boundary is almost saturated, so that the rate at which any route enlarges depends upon the amount of flow rather than upon dissolution kinetics (discharge-controlled stage). For a given conduit, enlargement rates increase slowly, as the discharge through the path is severely restricted by the narrower downstream parts. When breakthrough of L_{90} has occurred, enlargement accelerates dramatically and promotes a further large increase of discharge, so that a "run-away" condition develops. Enlargement rates in gypsum are further accelerated as turbulent conditions set in due to the mass transport control of dissolution kinetics. Those conduits that achieve early breakthrough are able to increase their discharge either by capturing water from neighbouring conduits or by extending their primary catchment. This situation emphasizes the importance of initial differences in hydraulic efficiency between fissures, and also explains the competitive early development of conduits.

However, this is already the stage when dissolution kinetics take control of enlargement. It is demonstrated (Palmer, 1984) that the enlargement rate of conduits in limestones does not increase in an unlimited way, but levels off at a maximum that is roughly of the order of 1mm per year. From then on, all successful conduits enlarge at almost identical rates. Such a limit has not been derived for gypsum by calculations, but it can be assumed that it is much higher than for limestone and is not achieved within reasonable flow rates. In most cases conduits in gypsum will continue to grow at accelerated rates until discharge is able to grow. For conduit growth in gypsum, the higher solubility of the rock and faster dissolution kinetics mean that the development stage in unconfined settings is much shorter than in the case of carbonate speleogenesis. The "runaway" development and competition of alternative flow paths under phreatic conditions are better expressed in gypsum than in limestones. This is the main reason for linear phreatic caves with poorly developed side passages being so typical within gypsum karst (see section 6.3).

When conduit enlargement is sufficient to carry more water than a catchment can deliver, the system switches to catchment control, and water table/vadose conditions are established (Palmer, 1984). In unconfined gypsum karst settings most active conduits adjust their sizes rapidly to accommodate the highest possible discharge (Forti, 1993), so that floods due to passage constrictions rarely occur in gypsum caves.

Confined conditions. The typical architecture to be considered comprises two "normal" (non-karstic) aquifers separated by a gypsum unit with fissure permeability, with the whole system confined by an upper, non-karstifiable, aquitard (Fig.3). Some vertical upward head gradient between the aquifers exists, and there is slow discharge from the system through the upper confining bed along structural weaknesses.

The head gradient between the lower and upper confined aquifers drives a slow flow through connected fissure paths. Initiation is slow, with wide tapers at the inputs to the fissures and abrupt terminations at the propagation front. When breakthrough occurs, the successful path increases its discharge to some extent, but initially not significantly. This is because flow through the system is governed (restricted) by the resistance of the upper confining bed, rather then by the available recharge, as in phreatic conditions. Increase in the path width after breakthrough locally minimizes the head gradient between the confined aquifers, so that the flow through it does not increase dramatically. It is then governed solely by the transmissivity of the upper confining bed, which is roughly constant and normally low. The fundamental difference between artesian and phreatic speleogenesis is that in the former case there is no dramatic boost in conduit enlargement rates to compare with that experienced under phreatic conditions (Fig. 4). The dynamics of conduit growth differ little between the initiation and development phases, and competitive development, common under phreatic conditions, is inhibited in artesian settings.

The development of conduits under confined conditions is rather slow and uniform, as the enlargement rates for all paths in the network will be essentially compatible and constant. The above consideration, which applies not only to gypsum, explains why maze cave patterns are such a common product of artesian speleogenesis. The last stage of artesian speleogenesis is terminated by localized breaching of hydrogeological confinement (such as along tectonic faults or by incising valleys), marked by a drastic increase of flow through the system. Flow will accelerate locally, causing increased enlargement rates along certain flow routes or zones, but this does not modify the already configured cave pattern significantly.

Major discrete tectonic fissures intersecting the whole gypsum bed represent a different development scheme (Fig.3-B). Their initial resistivity to flow is potentially low, if their width is large, so conduits grow fairly rapidly as breakthrough conditions might be present from the very begin-

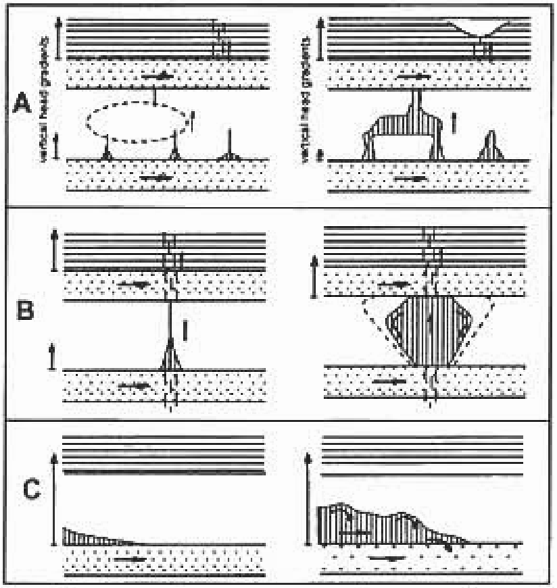

Fig. 3. Initiation (left) and development (right) of conduits and voids under artesian conditions. A = along common fissures; B = along a major tectonic fault; C = along a contact with an adjacent aquifer.

ning. However, enlargement rates do not accelerate significantly, as further increase of flow velocity is inhibited by the reduction in the head gradient between the upper and lower aquifers. As soon as an appreciable through space is created, much of the flow can reach the top of the gypsum without touching the gypsum walls. Under generally slow flow conditions, downward convection circulation cells develop, giving rise to an "inverted tapering" of the conduit shape (see more about the natural convection effect in the sub-chapter below). With further growth of through voids the head gradient between the adjacent aquifers tends towards zero, and circulation is driven largely by natural convection. This mechanism appears to be responsible for the continuing lateral growth of large voids in gypsum, and for triggering major breakdowns and the formation of vertical through structures (VTS) in coverbeds (see Chapter I.10 for details).

Dissolution also occurs along the contacts with the adjacent aquifers. In general, conduits initiate as tapers at inputs (Fig.3-C), however the actual mechanisms and shapes of the initiation and development depend also on the nature of the initial flow paths. These may follow bedding plane paths between solid rocks, be along the interface between solid gypsum and granular or porous aquifer material, or along water-conducting fissures in contact with the solid gypsum, and so on. Consequently, dissolved gypsum can be removed down-gradient along a bedding-related flow path, or in a direction normal to the lithological interface by diffusion or convection, then outflowing with the regional flow. The latter situation is most favourable for conduit initiation and development, as concentrations decrease at the interface, hence increasing the overall gypsum dissolution rate. Flow paths are guided by the arrangement of connected initial apertures along bedding planes, or by channels determined by intersections of fissures in adjacent insoluble rock with the bedding contact, or by more transmissive zones in adjacent granular material. When an aquifer underlies the gypsum, and some tapered space is created, further enlargement can be promoted by natural convection circulation. It is likely that some large voids can develop in this way.

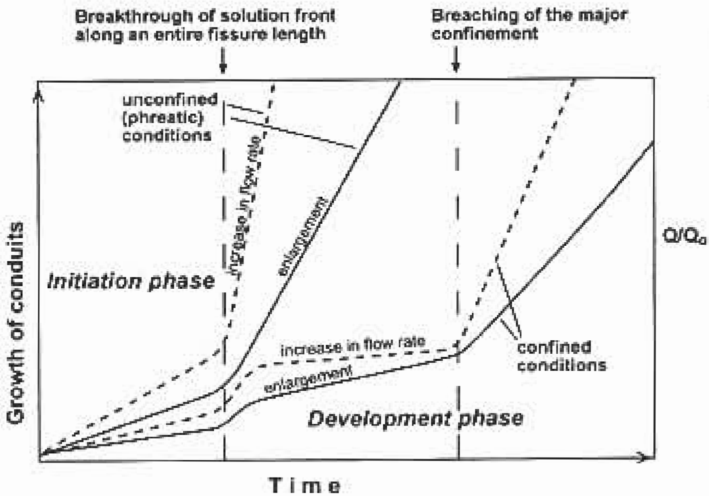

Fig. 4. Schematic diagram illustrating the dynamics of conduit growth through the initiation and development phases under phreatic and confined conditions.

5.3. Development at the water table and in the vadose zone

With the onset of the entrenched karst stage, vadose conditions become increasingly predominant, with the continued possible existence of water table and phreatic zones in the lower parts of massifs. Rapid enlargement of artesian and/or phreatic conduits occurs at the water table, particularly if annual fluctuations of major surface river levels cause periodic backflooding into a cave. In more stable conditions, such as in the interiors of watershed massifs, extensive horizontal notching may develop, promoted by water density stratification (see the sub-chapter below). In the vadose zone, cave development is concentrated along vertical percolation paths and free stream courses, but is very active locally. Hydrochemical data from different regions suggest that groundwaters in the vadose zone never attain saturation with respect to gypsum. More dispersed dissolutional enlargement may occur due to the action of condensation waters, but this effect tends to be localized in certain zones (see sub-chapter 5.4).

5.4. Speleogenetic effects of water density differences

Because dissolution always leads to solvent density increase, gravitational separation of water, and natural convection due to this effect, are inherently involved in, and affect, the cave development process. The effect may be significant in limestones (Curl, 1966) although is much more pronounced in gypsum (Kempe, 1972, 1975) and in salts (Frumkin, 1994) due to the higher solubility of these rocks. A recent overview and further elaboration of the issue has been provided by Klimchouk (1997b).

Water density difference effects generally become notable at the development stage, when substantial spaces are created by forced convection dissolution. In artesian settings the effects may also contribute to conduit initiation. When continuous or periodic recharge of fresh water occurs, dissolution of the rock sets up density gradients, which cause gravitational separation

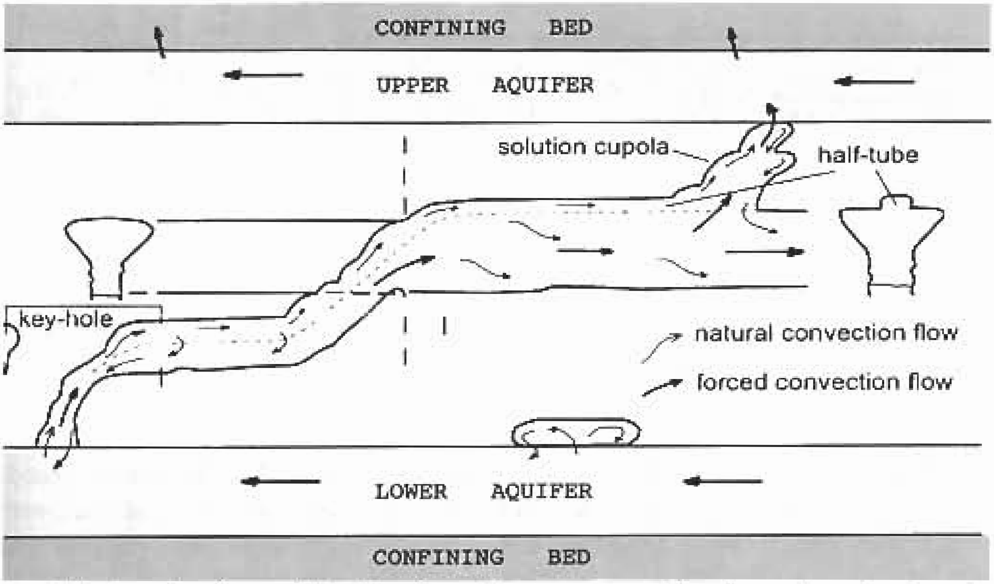

Fig. 5. The formation of upward dissolution forms by buoyant currents. The diagram shows the schematic relationship between lines of natural and forced convection flow on the mature stage of artesian speleogenesis, when conduit connection has already been established through the gypsum, but forced flow is slow due to the major constraint of the upper confining bed.

(stratification) of water and drive natural convection circulation. The phenomenon may operate at the local scale (e.g. in a cave lake) or at the scale of an aquifer (e.g. in artesian aquifers or across the water table zone).

Natural convection circulation and its speleogenetic effect are most pronounced in artesian settings because of sluggish flow conditions and low velocities, and also due to the commonly occurring recharge of gypsum from below. When sharp tapers are created at a fissure input at the base of a gypsum bed, dissolution is further promoted by natural convection circulation. After dissolving gypsum and increasing in density, part of the water returns downwards into the underlying aquifer and joins the regional flow output. It is quite possible that, under these conditions, dissolution driven by natural convection contributes even more to the upward propagation of enlargement through the fissure than does the penetration distance mechanism, driven by forced seepage, considered in sub-chapter 5.1. This view is supported by the common occurrence in many artesian caves of blind cupolas and domepits up to 10-15m high, with very tight fissures recognizable at their apices. In this way, vertical hydraulic connectivity between fissures arranged at different levels within a gypsum sequence is promoted, such that the effect facilitates the build-up of 3-D cave patterns. In more developed systems, where connection with the upper aquifer is established, directed (unlooped) buoyant currents operate, as less dense water always tends to occupy the uppermost available space. This is suggested (Klimchouk, 1997b) to account for the formation of at least some keyhole sections and ceiling half-tubes under artesian conditions (Fig. 5). In the case of large spa-

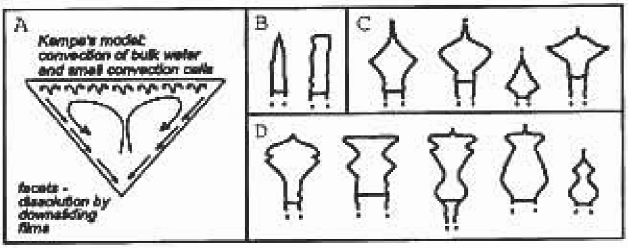

Fig. 6. Kempe's (1972) model of passage development by dissolution due to natural convection (A), and examples of typical cross-sections from the gypsum caves of the Western Ukraine showing varieties of notching or facetting effects.

ces that provide direct hydraulic connections between adjacent aquifers, downward convection is probably the main mechanism of lateral wall retreat (see Fig. 3-B and the sub-chapter above). Also, natural convection due to density differences is important to the development of cavities along the contacts between gypsum and underlying aquifers (Fig. 3-C).

Under shallow phreatic conditions, characteristic tip-down triangle cross-sections develop, with flat ceilings ("Laugdecke" in German), combined with inclined facets. They are quite common in many gypsum caves in Germany, the Western Ukraine, the Urals, Siberia and elsewhere, and have been studied in details and modelled theoretically by Kempe (1972) and Kempe et al. (1975). They are formed by dissolution in the uppermost, aggressive, layer of water, where patterns of small up- and downwelling convection cells ("salt-fingers") operate due to small local density differences (Fig. 6-A). Inclined facets are formed due to conduit-scale convection circulation, where films of water slide downwards along the walls, with progressive decrease in dissolutional potential.

Horizontal notching caused by chemical stratification of water, with the highest dissolution rates in the uppermost layer, may be a common morphological effect in caves within all major karstifiable lithologies (Ford & Williams, 1989).,It is, however, best displayed in salts (Frumkin, 1994) and gypsum (Klimchouk & Aksem, 1988; Klimchouk, 1997b; see also Fig. 6-C, 6-D).

5.4. Speleogenetic effects of condensation dissolution

In the aerated zone of well-karstified entrenched or exposed karst massifs, condensation processes can make a significant contribution to groundwater recharge. The role of condensation in karst hydrogeology and speleogenesis has been well studied in the Soviet Union, where Lukin (1962, 1969), Dubljansky (1970) and Dubljansky & Sotzkova (1982) elaborated theoretical issues and provided assessments and reviews of available field data, including some from gypsum karst areas. Cigna & Forti (1986), Forti (1991, 1993) and Calaforra et al. (1992, 1993) addressed the issue with particular regard to gypsum caves, but they were unaware of previous Soviet studies.

The amount of condensation water that can be formed in caves and fissures depends upon climate, the intensity of air exchange and temperature differences between the outside and in-cave atmospheres. It is most pronounced during warm seasons, under temperate climatic conditions and, especially, in the semi-arid zone. Water that condenses in transitional micro-climatic zones in caves is very aggressive, and causes substantial dissolution. The role of condensation corrosion in cave development is more important in gypsum than in carbonates, due to gypsum's high solubility and fast dissolution kinetics.

Lukin (1969) estimated that every cubic metre of air leaves 10g of water condensed on rock surfaces while passing throughout Kungurskaya Cave (in the Pre-Urals) during the warm season. Calaforra et al. (1993) suggested that, in the Cueva del Agua area of the semi-arid karst of Sorbas, all of the perennial base flow (about 1 L/s comprising 25% of the total discharge of the aquifer) was provided by condensation processes active inside the cave. Forti (1993) estimated that condensation accounts for more than 60% of the recharge of the karst aquifer associated with the Cueva del Leon in Argentina. Estimations by Dubljansky for different regions gave seasonal (the warm season) rates of condensation generated flow that vary from 1.4 to 9.7 L s^{-1} km^{-2}, comprising from 5.9 to 85% of the total recharge (precipitation minus evapo-transpiration). Klimchouk et al. (1988) approximated the rates of gypsum dissolution caused by condensation in the local zone inside the Optimisticheskaja Cave in the Western Ukraine to vary from -0.001 to -0.005 mg cm^{-2} day-1 according to season, with actual values reaching up to -0.02 mg cm^{-2} day^{-1} during certain 1-2 month periods. Thus, dissolution due to condensation can be a notably active agent of cave development.

6. Types of caves in gypsum karst

The above consideration of cave origin and development in gypsum provides a guide to the genetic classification of gypsum caves presented in Table 2. Below, the main cave types are briefly characterized, with reference to representative examples.

6.1. Discrete voids

Caves of this type develop commonly under artesian conditions, where the gypsum is underlain by an aquifer, and its own initial permeability is either very inhomogeneous, is determined by discrete major tectonic fissures, or is negligible. Their origin and development mechanisms are described in the section above (see also Chapter II.5). The best documented examples are caves in the Sangerhausen and Mansfeld districts of Germany, encountered through the centuries in the course of mining operations at depths up to 400m at the base of the Zechstein gypsum (Kempe, this volume). They are large voids, commonly isometric, or elongated along the major tectonic fissures like the Wimmelburger Schlotten (see Figs. 2-D and 4 in Chapter II.5). About 100 cavities of this type are known in the region. Natural convection circulation, driven by water density gradients, with dissolved gypsum flowing out with the regional flow in the underlying aquifer, is believed to play an important role in the development of such cavities (Kempe, this volume).

Breakdown of large voids formed in such a way is probably the main trigger of the development of vertical through structures (VTS). The latter is the generic term suggested for features including breccia pipes, collapse columns, and so on, common in deep-seated intrastratal karsts (see Chapter I.9 for details).

6.2. Maze caves

Maze caves constitute about 21% of the 197 largest recorded gypsum caves in the world, but their proportion increases to 41% among caves over 1000m long, and to 54% of gypsum caves

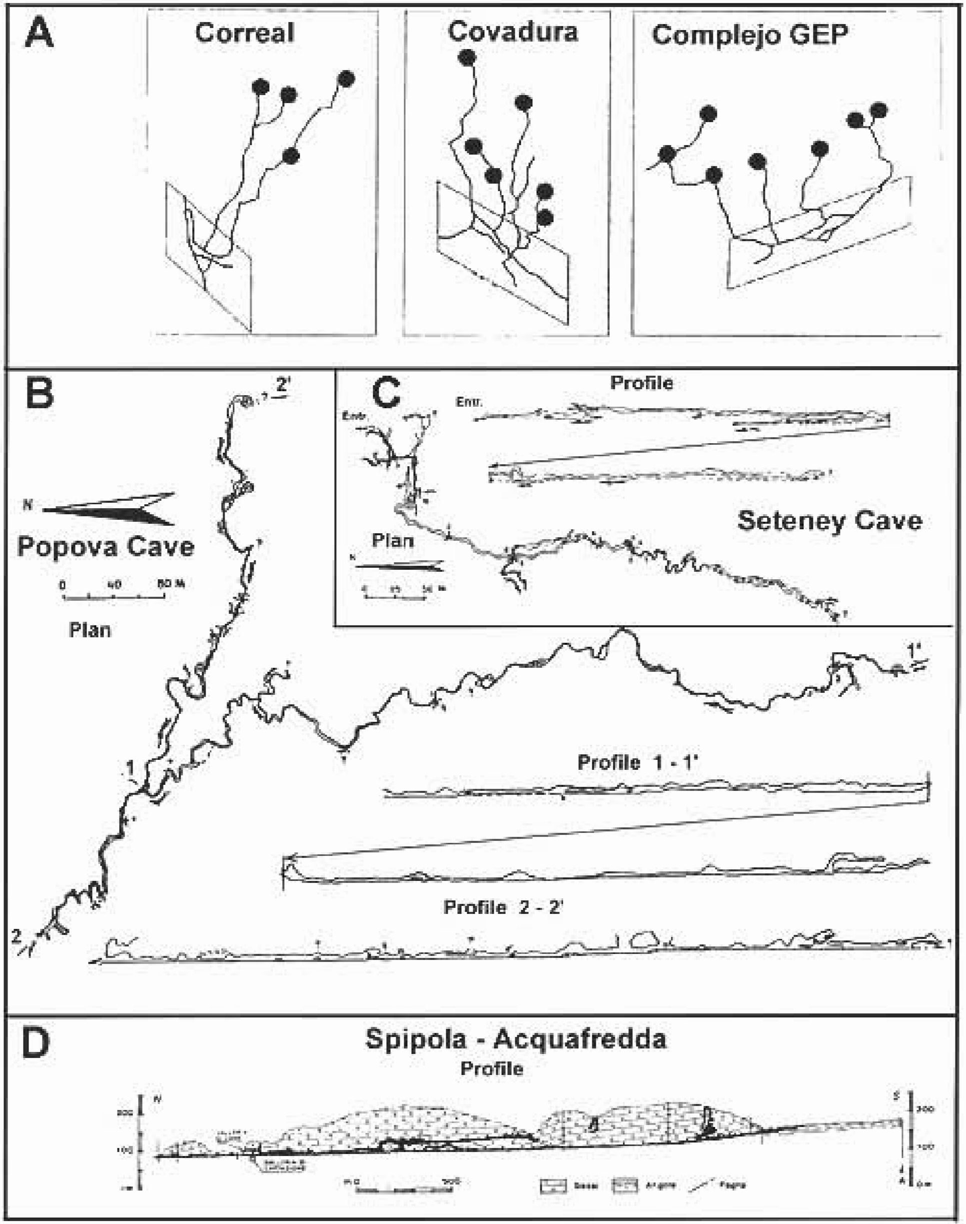

Fig. 7. Generalised sketches, plans and profiles of some gypsum "through caves": A = in Sorbas, Spain. Feeding dolines are indicated by dots. (After Calaforra, 1996); B and C = in the Ekeptze-Gadyk massif, North Caucasus, Russia (after Ostapenko, 1994); D = the Spipola-Acquafredda system in the Emilia Romagna, Italy (After Grimandi, 1987).

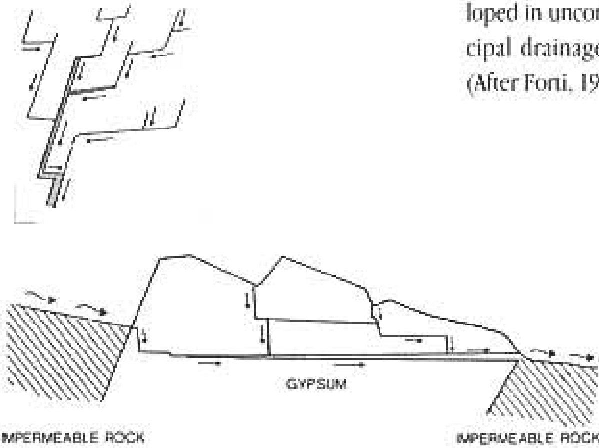

Fig. 8. Sketch of a typical gypsum karst system developed in unconfined conditions, consisting of a principal drainage tube with few and short effluents (After Forti, 1993).

points. They are collectively termed "through caves" in this account. As is shown in section 5.2, the "run-away" development and competition of alternative flow paths in unconfined conditions is exaggerated in gypsum, so that normally only one passage develops between input and output points (Fig. 7). When there are multiple sink points, a dendritic pattern may develop, as minor flow paths will ultimately connect to the nearest major successful conduit that serves as a drain (Fig. 8). Speleogenesis in gypsum under unconfined conditions creates extreme anisotropy of permeability, with rather simple and strongly hierarchical networks. Forti (1993) claimed the latter to be the principal characteristic of speleogenesis in gypsum, referring to confined speleogenetic environments as rare special cases. The actual situation can be said to be the direct opposite, considering that exposed or entrenched gypsum karsts, with no inherited caves at all, comprise only a minor part of the gypsum formations undergoing karstification (see also Chapter I.4).

Because of the fast development stage, "through caves" in gypsum adjust rapidly to the present base level. They also commonly develop along intercalated insoluble and poorly permeable (if compared with the now karstified gypsum) layers, or along the top of the basal formation, perched within the vadose zone. Perched streams "drop" into the nearest major tectonic fissure, forming vadose pits that connect different levels (Fig. 9). Intercalations in gypsum sequences play an important role in the early development of conduits under phreatic conditions. During the vadose stage, erosion of insoluble passage floors may become the predominant mechanism of their further development. This feature is well illustrated in the caves of Sorbas (Fig. 10).

"Through caves" are common in almost every entrenched and denuded gypsum karst area. The most representative and best documented examples are in the Emilia-Romagna and Sicily regions of Italy, the Belomorsko-Kulojsky, Pre-Urals and North Caucasus regions of Russia, at Sorbas in Spain, and in New Mexico and Oklahoma in the United States.

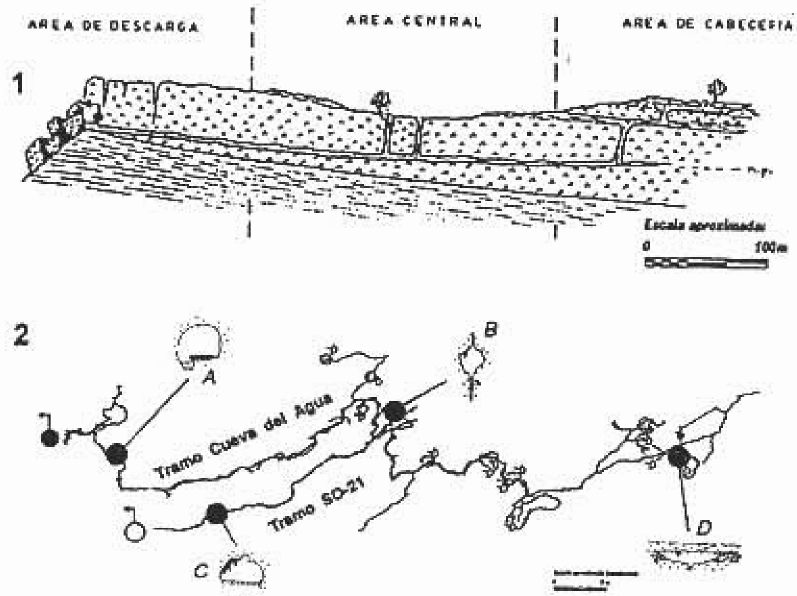

Fig. 9. Schematic profile (1) and plan (2) of the Cueva del Agua system, a typical "through cave" developed in the denuded exposed karst of Sorbas, Spain. Cross-sections of passages differentiate throughout the system: A = active passages in the downstream section, B = relict (abandoned) passages, C = phreatic passages with superimposed vadose canyons in the central section, D = vadose passages in the upstream section, where they developed by erosion of clastic intercalation layers (Adopted from Calaforra, 1996).

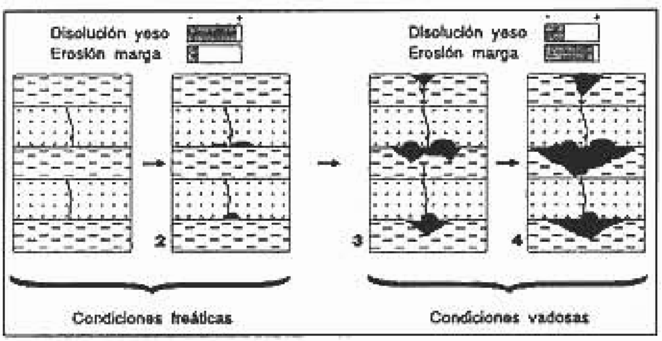

Fig. 10. Speleogenesis of interstratal passages in the gypsum karst of Sorbas, Spain. 1-2 = initiation and development of solutional conduits in phreatic conditions above the contact with marl intercalations, 3 - 4 = development of passages in vadose conditions by erosion of clastic intercalation layers (After Calaforra, 1996).

6.4. Vertical pipes

Vertical dissolution pipes, also known as organ pipes, or "komins" in the Russian literature, represent a very common feature of entrenched intrastratal karst. They develop downwards from a suitable protective bed at the top of the gypsum (commonly limestone or dolomite), due to focused dissolution by groundwater that percolates through the overburden, or leaks from perched aquifers above the gypsum (see Fig. 3 in Chapter I.10). Pipes cut across the whole gypsiferous stratum, or down to the water table, commonly intersecting relict lateral caves. Their density in a given area depends mainly upon the abundance of percolating trickles in the coverbeds, and can be very high in some places, perhaps up to several hundred per km2. Pipe diameter depends upon the amount of percolating water. New pipes develop quite rapidly, reaching a diameter of 1 to 3m, before their growth rate slows down. Rapid growth of new pipes is evidenced by an example 1m in diameter in Zolushka Cave in the Western Ukraine, which developed during 35 years after a borehole drilled from the surface caused a new leakage point from the perched aquifer above through the intervening clay. Dissolution pipes commonly induce breakdowns and vertical through structures (VTS) in coverbeds (see Chapter I.10). They also provide foci for doline development where coverbeds are scoured by denudation.

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