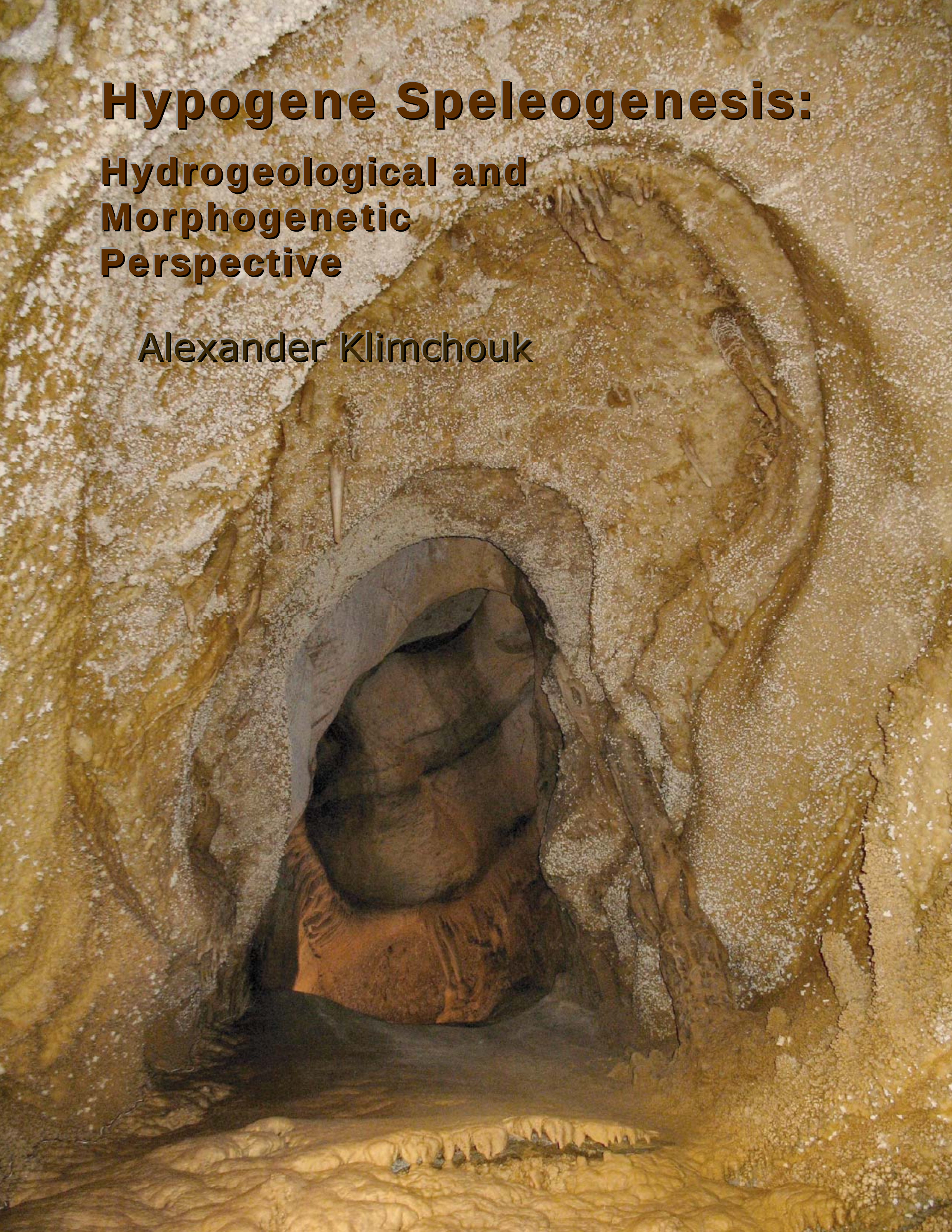
Hypogene Speleogenesis:

Hydrogeological and Morphogenetic Perspective

Alexander Klimchouk



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Dedication

To Kimberly (Kim) Cunningham, an outstanding person, a genuine friend and a bright cave scientist, who left an indelible impression on people who where lucky to know him.

Cover photo:
Front: A rising chain of cupolas in Caverns of Sonora, TX, USA (Photo by A. Klimchouk)
Back: A dome possibly leading to a higher story of passages (an exploring caver climbing a rope provides a scale). See Plate 11 for a broader view. Echo Chamber in Lechuguilla Cave, NM, USA (Photo by S. Allison).

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Foreword

In 1998, the National Cave and Karst Research Institute (NCKRI) was established by a U.S. Congressional mandate to facilitate and support cave and karst research, stewardship, and education. I am delighted to introduce this new publication series, NCKRI Special Papers, as an essential part of NCKRI's efforts to meet that mandate. I'm equally pleased that this first book in the series is the highly important work of NCKRI's first visiting scholar, Dr. Alexander B. Klimchouk.

Caves are resources hidden from the view of the general public and most scientists. Their value often goes unrecognized because they are either not seen or misunderstood. Historically, caves were ignored by many geoscientists, in part because they did not "follow the rules" of groundwater behavior and thus "had" to be anomalous features of little significance. While this view has mostly changed, many scientists who realized the significance of caves had and still have the mistaken notion that areas of carbonate and evaporite rocks that contain few or no caves are not karst. This book shatters those myths and makes great strides in explaining what had been some of the most puzzling aspects of karst hydrogeology.

Dr. Klimchouk carefully explains the origin of hypogenic caves and karst, and demonstrates it with a rich, international array of examples and data. While most karst literature focuses on epigenic karst, formed by descending groundwater, hypogenic karst stems from ascending groundwater. Understanding the characteristic set of hypogenic morphological and hydrological features, and the processes that create them, is crucial for developing accurate models, and effective management plans for these karst systems. This is vital because hypogenic karst is especially poorly expressed at the surface, and so its vulnerability as a public water supply, risk of sinkhole collapse, and value as a mineral resource can be severely underestimated.

While this book focuses on karst hydrogeology and speleogenesis, it also has important implications for many other disciplines, such as understanding the ranges and speciation of cavernicolous organisms, landscape evolution, and the distribution of paleontological and archeological deposits, to name a few. At a fundamentally crucial level, the great geographic breadth of hypogenic karst will soon be realized directly as a result of this work. Certainly, some concepts presented here will be refined with continued research, but this book firmly establishes a new paradigm that will guide much karst research for decades to come.

Dr. George Veni
Executive Director, NCKRI

April 2007

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Abstract

This book provides an overview of the principal environments, main processes and manifestations of hypogenic speleogenesis, and refines the relevant conceptual framework. It consolidates the notion of hypogenic karst as one of the two major types of karst systems (the other being epigenetic karst). Karst is viewed in the context of regional groundwater flow systems, which provide the systematic transport and distribution mechanisms needed to produce and maintain the disequilibrium conditions necessary for speleogenesis. Hypogenic and epigenic karst systems are regularly associated with different types, patterns and segments of flow systems, characterized by distinct hydrokinetic, chemical and thermal conditions. Epigenic karst systems are predominantly local systems, and/or parts of recharge segments of intermediate and regional systems. Hypogenic karst is associated with discharge regimes of regional or intermediate flow systems.

Various styles of hypogenic caves that were previously considered unrelated, specific either to certain lithologies or chemical mechanisms are shown to share common hydrogeologic genetic backgrounds. In contrast to the currently predominant view of hypogenic speleogenesis as a specific geochemical phenomenon, the broad hydrogeological approach is adopted in this book. Hypogenic speleogenesis is defined with reference to the source of fluid recharge to the cave-forming zone, and type of flow system. It is shown that confined settings are the principal hydrogeologic environment for hypogenic speleogenesis. However, there is a general evolutionary trend for hypogenic karst systems to lose their confinement due to uplift and denudation and due to their own expansion. Confined hypogenic caves may experience substantial modification or be partially or largely overprinted under subsequent unconfined (vadose) stages, either by epigenic processes or continuing unconfined hypogenic processes, especially when H_2S dissolution mechanisms are involved.

Hypogenic confined systems evolve to facilitate cross-formational hydraulic communication between common aquifers, or between laterally transmissive beds in heterogeneous soluble formations, across cave-forming zones. The latter originally represented low-permeability, separating units supporting vertical rather than lateral flow. Layered heterogeneity in permeability and breaches in connectivity between different fracture porosity structures across soluble formations are important controls over the spatial organization of evolving ascending hypogenic cave systems. Transverse hydraulic communication across lithological and porosity system boundaries, which commonly coincide with major contrasts in water chemistry, gas composition and temperature, is potent enough to drive various disequilibrium and reaction dissolution mechanisms. Hypogenic speleogenesis may operate in both carbonates and evaporites, but also in some clastic rocks with soluble cement. Its main characteristic is the lack of genetic relationship with groundwater recharge from the overlying or immediately adjacent surface. It may not be manifest at the surface at all, receiving some expression only during later stages of uplift and denudation. In many instances, hypogenic speleogenesis is largely climate-independent.

There is a specific hydrogeologic mechanism inherent in hypogenic transverse speleogenesis (restricted input/output) that suppresses the positive flow-dissolution feedback and speleogenetic competition in an initial flowpath network. This accounts for the development of more pervasive channeling and maze patterns in confined settings where appropriate structural prerequisites exist. As forced-flow regimes in confined settings are commonly sluggish, buoyancy dissolution driven by

either solute or thermal density differences is important in hypogenic speleogenesis.

In identifying hypogenic caves, the primary criteria are morphological (patterns and meso-morphology) and hydrogeological (hydrostratigraphic position and recharge/flow pattern viewed from the perspective of the evolution of a regional groundwater flow system). Elementary patterns typical for hypogenic caves are network mazes, spongework mazes, irregular chambers and isolated passages or crude passage clusters. They often combine to form composite patterns and complex 3-D structures. Hypogenic caves are identified in various geological and tectonic settings, and in various lithologies. Despite these variations, resultant caves demonstrate a remarkable similarity in cave patterns and meso-morphology, which strongly suggests that the hydrogeologic settings were broadly identical in their formation. Presence of the characteristic morphologic suites of rising flow with buoyancy components is one of the most decisive criteria for identifying hypogenic speleogenesis, which is much more widespread than was previously presumed. Hypogenic caves include many of the largest, by integrated length and by volume, documented caves in the world.

The refined conceptual framework of hypogenic speleogenesis has broad implications in applied fields and promises to create a greater demand for karst and cave expertise by practicing hydrogeology, geological engineering, economic geology, and mineral resource industries. Any generalization of the hydrogeology of karst aquifers, as well as approaches to practical issues and resource prospecting in karst regions, should take into account the different nature and characteristics of hypogenic and epigenic karst systems. Hydraulic properties of karst aquifers, evolved in response to hypogenic speleogenesis, are characteristically different from epigenic karst aquifers. In hypogenic systems, cave porosity is roughly an order of magnitude greater, and areal coverage of caves is five times greater than in epigenic karst systems. Hypogenic speleogenesis commonly results in more isotropic conduit permeability pervasively distributed within highly karstified areas measuring up to several square kilometers. Although being vertically and laterally integrated throughout conduit clusters, hypogenic systems, however, do not transmit flow laterally for considerable distances. Hypogenic speleogenesis can affect regional subsurface fluid flow by greatly enhancing initially available cross-formational permeability structures, providing higher local vertical hydraulic connections between lateral stratiform pathways for groundwater flow, and creating discharge segments of flow systems, the areas of low-fluid potential recognizable at the regional scale. Discharge of artesian karst springs, which are modern outlets of hypogenic karst systems, is often very large and steady, being moderated by the high karstic storage developed in the karstified zones and by the hydraulic capacity of an entire artesian system. Hypogenic speleogenesis plays an important role in conditioning related processes such as hydrothermal mineralization, diagenesis, and hydrocarbon transport and entrapment.

An appreciation of the wide occurrence of hypogenic karst systems, marked specifics in their origin, development and characteristics, and their scientific and practical importance, calls for revisiting and expanding the current predominantly epigenic paradigm of karst and cave science.

Introduction

Most studies of karst systems are concerned with shallow, unconfined geologic settings, supposing that the karstification is ultimately related to the Earth's surface and that dissolution is driven by downward meteoric water recharge. Such systems are epigenic (hypergenic). Concepts and theories developed for unconfined karst systems overwhelmingly predominate in karst and cave science, particularly in karst hydrology and geomorphology, forming a core of the current karst paradigm. Hypogenic karst, originating from depth and not related to recharge from the overlying surface, although becoming more recognized during the last two decades, remains poorly understood and integrated into the bulk of karst science.

There are specific reasons for this bias, arising from the historic paths through which the knowledge of the karst domain evolved. Epigenic karst systems evolve when soluble rocks occur in the shallow subsurface or become exposed, so they inherently express surface components, readily available for observations and affecting many aspects of human activity. Epigenic karst systems form by water infiltrating or in-flowing from overlying or immediately adjacent recharge surfaces and develop in genetic relation to landscape. Caves commonly have a hydrologic connection with the surface and "genetically inherent" entrances. Karst knowledge in Western Europe and North America had originally commenced mainly from exploration and study of such caves. These factors in combination led to a deeply rooted belief that epigenic unconfined karst systems overwhelmingly predominate[1]. Karst features, routinely encountered by wells and mines in soluble rocks at substantial depths, were (and still are) commonly regarded as paleokarst features, originally formed in epigenic settings and subsequently buried under younger sediments.

Some explored caves, however, display patterns, morphologies, sediments, and minerals that do not readily conform to established concepts of epigenic karst development and speleogenesis. Until recently, they were (and in many cases still are) explained in terms of epigenic/unconfined speleogenesis, which led to numerous more or less obvious misconceptions and controversies. Over the last 20 years there has been a rapid increase in the development of speleogenetic ideas implying a hypogenic and/or confined origin of caves, with reference to a deep source of acidity or to a confined flow system. However, in the general context of the predominant karst paradigm, such caves are still largely regarded as special, aberrant cases. In his classic work on cave origin, Palmer (1991) estimated that hypogene cave systems account for only about 10% of the studied cave systems, although they include some of the largest ones. Since then, ongoing re-interpretation of some known caves has probably increased this percentage. Enhanced understanding of hypogenic speleogenesis and the refinement of criteria for their recognition are going to further increase this figure. More important is the fact that hypogene/confined karst systems are globally much more widespread than it is now believed, and the relatively small fraction of known caves of this type is merely an exploration bias resulting from their genetic irrelevance to the surface and inherent lack of accessibility.

[1] In contrast, in some regions where karstology as a scientific discipline preceded cave exploration, and where "covered" (deep-seated) karst settings are widespread, such as in the former Soviet Union, deep-seated, hypogene, confined karst and some relevant processes have been long recognized, at least in general terms.

Significant advances in understanding of speleogenesis in hypogene (deep-seated) and confined (artesian) settings made during recent years remain somewhat fragmented and uncoordinated. This is partly because discussions of the particular cases of "atypical" speleogenesis (sulfuric acid, hydrothermal, in some sense – speleogenesis in evaporites) focus attention on geochemical processes of solutional porosity creation with the hydrogeologic framework of cave formation often remaining poorly understood. There is a misleading trend to label particular types of speleogenesis, or even types of karst, by the dissolutional mechanism assumed to have created the caves. This obscures the fact that most hypogenic/confined karst systems share many major common features in their geo/hydrogeological settings, patterns and morphologies. Although geochemical attributes and dissolution mechanisms are indispensable components of the speleogenetic environment, the principal component is groundwater flow. Other attributes largely depend on the position of a given karst system in the basinal groundwater flow system and evolution of boundary conditions. By way of analogy, in creating dissolutional porosity the groundwater flow system is a "master," the type of recharge is a "tool" and dissolutional mechanisms are the "fuels" to power the tools. The shape, pattern and size of holes produced by a tool (dentist drill, hand drill, borehole drill bit, bulldozer or excavator) depend more on the intentions of a master and the type of tool rather than on the fuel that drives it.

This paper intends to give an overview of the principal environments, main processes and manifestations of hypogenic speleogenesis, in order to show the place of hypogenic karst systems in the basinal groundwater flow systems, thus demonstrating the common genetic background of various styles of hypogenic karst and caves that were previously considered unrelated, specific either to lithologies and/or chemical mechanisms. I intend to demonstrate the fundamental importance of the type of flow system in the formation of hypogenic (confined) karst and caves, and establish hypogenic karst as one of two major types of karst systems.

The appreciation of the widespread occurrence of hypogenic karst systems, marked specifics in their origin and development, and their scientific and practical importance, calls for revisiting and expanding the current paradigm of karst and cave science.

1. Basic concepts and terminology

1.1 Karst and speleogenesis

As this book focuses on phenomena and processes poorly integrated into the established conceptual framework of karst, it is necessary to clarify some basic concepts and terminology.

Most modern texts, encyclopedias, and reviews define karst from a largely geomorphological perspective: *"Karst is terrain with distinctive hydrology and landforms arising from the combination of high rock solubility and well-developed solutional channel (secondary) porosity underground"* (Ford and Williams, 1989; Ford, 2004).. Distinctive landforms and surface hydrology, however, are not necessarily characteristic for hypogenic karst.

Although it is often claimed that approaches, concepts and methodologies to study karst differ between geomorphology and hydrology, the modern conceptual framework in karst hydrology seems to remain constrained by the historically prevailing largely geomorphological paradigm of karst as an epigenic unconfined system, closely related to surface drainage (White, 2002; Bakalowicz, 2005). Moreover, earlier firmly rooted, historically biased views that only karst features in carbonates are considered a true karst are still reiterated in modern publications:

"Karst features mainly occur in carbonate rocks, limestone and dolomite, in which formations it is considered as true karst" (Bakalowicz, 2005);

"Karst: Landforms that have been modified by dissolution of soluble rocks (limestone and dolostone)" (Poucher and Copeland, 2006);

"Karst is defined as a limestone landscape with underground drainage" (Waltham, in Luhr, 2003).

Another illustration of the poor integration of hypogenic karst into modern karst knowledge is that the recent two fundamental encyclopedias on caves and karst (Gunn, 2004; Culver and White, 2004) do not contain distinct entries on hypogenic karst, although they do consider many aspects of hypogenic karst and certainly hypogenic caves (as they include some of the largest and most important caves in the world).

Although such focusing was certainly productive in consolidating the conceptual framework and methodology in studying the epigenic type of karst systems, the situation with the predominantly epigenetic approach to karst, reflected in general reviews on the subject, hinders progress in recognition and study of hypogenic karst.

An emerging approach to karst hydrogeology is more integrative and universal by encompassing the whole range of karst processes and phenomena. Following proposals of Huntoon (1995) and Klimchouk and Ford (2000), karst is defined here as *"an integrated mass-transfer system in soluble rocks with a permeability structure dominated by conduits dissolved from the rock and organized to facilitate the circulation of fluids."* Whether karst is expressed at the surface or not is irrelevant. A karst system can operate in the subsurface without any apparent relationship to the surface, being represented exclusively by underground forms that draw their input water from and discharge their output water to other non-karstic rocks.

Speleogenesis is viewed as *"the creation and evolution of organized permeability structures in a rock that have evolved as the result of dissolutional enlargement of an earlier porosity"* (Klimchouk and Ford, 2000, p.47), making it the most essential part of the karst concept. One can assert that karst is a function of speleogenesis, a statement for which the validity is particularly evident in cases where the surface landscape component is absent or subdued as in hypogenic karst. The notion of "karst," however, is broader than that of "speleogenesis," as it includes features and phenomena induced by speleogenesis but not encompassed by it.

Though some uncertainties still remain in the scope of the related and overlapping concepts/terms (discussed in the next section), three basic genetic settings are broadly recognized now for caves (Ford and Williams, 1989; 2007; Klimchouk *et al.*, 2000; Ford, 2006): 1) *coastal and oceanic*, in rocks of high matrix porosity and permeability; 2) *hypogenic*, predominantly confined, where water enters the soluble formation from below, and 3) *hypergenic (epigenic)*, unconfined, where water is recharged from the overlying surface. Although coastal and oceanic settings are commonly characterized by unconfined circulation, they are treated separately because of the specific conditions for speleogenesis determined by the dissolution of porous, poorly indurated carbonates by mixing of waters of contrasting chemistry at the halocline.

1.2 Hypogenic, confined and deep-seated speleogenesis

Hypogenic (or hypogene) caves are usually considered the opposite to the common *epigenic* caves formed by water recharged from the overlying or immediately adjacent surface due to carbonic acid dissolution. A more appropriate antonym to "hypogenic" is *hypergenic (or hypergene)*; the term widely used in Eastern Europe to denote processes operating near the surface through the action of descending solutions.

The term and concept of hypogenic speleogenesis has seen increasing use during the recent two decades, although still with some uncertainty in its meaning. Two approaches appear in recent works. Ford and Williams (1989) and Worthington and Ford (1995) defined hypogenic caves as those formed by hydrothermal waters or by waters containing hydrogen sulfide. Hill (2000a) tends to narrow the notion of hypogenic karst and speleogenesis to H_2S-related processes and forms. Palmer (1991) defined hypogenic caves more broadly, as those formed by acids of deep-seated origin, or epigenic acids rejuvenated by deep-seated processes. Later on, Palmer (2000a), presented the definition in a slightly modified, even broader form: *hypogenic caves are formed by water in which the aggressiveness has been produced at depth beneath the surface, independent of surface or soil CO_2 or other near-surface acid sources.* This modification is important, as it formally allows us to include in the class of features formed by still surface-independent but non-acidic sources of aggressiveness (such as aggressiveness of water with respect to evaporites). Reference to acid sources seems to be confusing however, as it again tacitly implies that features formed by non-acid dissolution are not pertinent.

Palmer's definition relies on the source of aggressiveness. The aggressiveness is a transient attribute of water, which can be delivered from depth or acquired within a given soluble formation (due to mixing or redox processes, for instance). It is suggested here that the definition of hypogenic speleogenesis should rather refer to the source of groundwater, as it is a medium of transport of aqueous and nonaqueous matter and energy, a reactive agent and a major component of the speleogenetic environment. Hypogenic speleogenesis is defined here, following the recent suggestion of Ford (2006), as "*the formation of caves by water that recharges the soluble formation from below, driven by hydrostatic pressure or other sources of energy, independent of recharge from the overlying or immediately adjacent surface*."

Hypogenic speleogenesis does not rely exclusively on certain dissolutional mechanisms; a number of dissolutional processes and sources of aggressiveness can be involved (see Section 3.6 below). Its main characteristic is the lack of genetic relationship with groundwater recharge from the overlying surface. In many instances, hypogenic speleogenesis is climate-independent. It may not be manifested at the surface at all (deep-seated karst). Hypogenic caves commonly come into interaction with the surface as relict features, largely decoupled from their formational environment, when ongoing uplift and denudation shift them into the shallow subsurface.

The concept of hypogenic speleogenesis is closely related to the notion of *artesian* or *confined* speleogenesis. These terms refer to the important aquifer condition, where groundwater is under pressure in a bed or stratum confined by a less permeable rock or sediment above it. The criterion of non-relevance of hypogenic speleogenesis to overlying surface recharge and sources of aggressiveness implies substantial separation of groundwater circulation from the overlying surface, *i.e.* some degree of confinement or rising flow. Groundwater rises through soluble but initially poorly permeable or heterogeneous formations. Most hypogenic speleogenesis initially occurs under confined conditions, which accounts, as will be shown in the following sections, for its most essential features (see also Klimchouk, 2000a, 2003a, 2003b, 2004). However, there is an evolutionary trend for hypogenic karst systems to lose their confinement in the course of uplift and denudation, and due to their own expansion. Hypogenic development may continue in unconfined settings, but confined conditions are the most essential element of hypogenic speleogenesis.

Other concepts that are relevant to hypogenic speleogenesis are *intrastratal karst* and *deep-seated karst*. *Intrastratal karst* is developed within rocks already buried by younger strata, where karstification is later than deposition of the cover rocks (Quinlan, 1978; Palmer and Palmer, 1989). This meaning does not relate to genesis but implies stratigraphic, although not necessarily hydrogeologic, separation of karst development from the surface by overlying non-soluble strata and emphasizes the evolutionary aspect (*i.e.* karst is later than the cover). Intrastratal karst can be deep-seated, subjacent, entrenched or denuded (the latter represents the former intrastratal karst); this subdivision has an evolutionary meaning and relates karst settings to the depth of erosional entrenchment and the degree of denudational exposure (Klimchouk, 1996a; Klimchouk and Ford, 2000). Depending on the cover lithology and the depth of erosional entrenchment, intrastratal karst can develop in confined or unconfined conditions. *Deep-seated karst* is developed without any exposure of the soluble rock to the overlying surface, and is not related to it. It implies ongoing contemporary karstification, so it should be differentiated from buried karst, or paleokarst. Deep-seated karst is always intrastratal, although the opposite is not always true. Deep-seated karst is predominantly confined and hypogenic; this is discussed further in the next section.

2. Karst in the context of the systematized and hierarchical nature of regional groundwater flow

Artesian basins are principal hydrogeologic structures at regional scales in predominantly layered sedimentary rocks (sedimentary basins) that contain stratiform groundwater bodies (layered aquifers); and *hydrogeological massifs* are tectonic block-faulted groundwater bodies with an overwhelming dominance of crosscutting fissure-conduit permeability (Zaitzev and Tolstikhin, 1971; Pinneker, 1977). Transitional types include *disrupted basins* and *layered massifs*. In cratonic regions and their passive margins, large artesian basins predominate, with subordinated hydrogeological massifs. Folded orogenic regions are characterized by the dominance of hydrogeological massifs, although small artesian basins are also common. Basins and massifs are commonly hydraulically connected, with massifs playing the role of marginal recharge areas.

Broad understanding of karst processes as a geological agent, one of the most powerful and universal illustrations of groundwater as a geological agent, is based on the growing recognition in mainstream hydrogeology of hydraulic continuity, the systematized nature and hierarchical organization of regional flow, and the great role of cross-formational communication in multiple-aquifer (multi-story) confined systems (e.g. Pinneker, 1982; Sharp and Kyle, 1988; Shestopalov, 1981, 1989; Tóth, 1995, 1999). Principal categories of karst-forming environments and resultant karst/ speleogenetic styles can be adequately understood and classified only within the context of regional groundwater flow systems, as they are regularly associated with distinct segments and evolutionary states of these systems. The works of Tóth (1995, 1999) provide a particularly useful and inspiring synopsis of the nature of the system, hierarchical organization, and the geologic role of regional groundwater flow systems.

Speleogenesis, like other natural effects produced by groundwater flow systems, is a result of interaction between groundwater and its environment, driven by the various components and attributes of the two respective systems seeking equilibration (Tóth, 1999). To cause speleogenetic development, dissolution effects of disequilibria have to accumulate over sufficiently long periods of time and/or to concentrate within relatively small rock volumes or areas. The systematic transport and distribution mechanism capable of producing and maintaining the required disequilibrium conditions is the groundwater flow system (Tóth, 1999). This is the single fundamental reason why the principal categories of karst and speleogenetic environments should be distinguished primarily on the basis of hydrogeologic considerations, rather than by the particular dissolutional mechanisms involved.

The development of groundwater circulation is broadly cyclic. The hydrogeologic cycle begins with marine sedimentation that is succeeded by tectonic subsidence and the formation of connate waters. It then encompasses uplift, with denudation and progressive invasion of meteoric waters into the reservoir. It may include the intrusion of magma with release of juvenile waters. It closes with a new marine transgression.

Groundwater circulation in a basin adjusts to the pattern of maximum and minimum fluid potentials. Large-scale groundwater flow in sedimentary basins can be driven by several forces, such as sediment compaction due to burial or tectonic compression, dehydration of minerals, continental landscape topography gradients, and density gradients due to temperature or solute variations. Following uplift and establishment of the continental regime and topography, gravity-driven flow systems of meteoric groundwater increasingly flush out connate and resurgent waters from a basin, although compaction-driven flow systems may still predominate in deep parts of basins. The basinal groundwater system may be even more complicated, heterogeneous and heterochronous, when a basin goes through two or more hydrogeologic cycles, and/or where a basin is deformed and subdivided by differential tectonic movements and/or intruded with magma, with sub-systems of different magnitude and origin, superimposed and mingled (Pinneker, 1982; Tóth, 1995).

Most known karst systems develop in continental domains dominated by gravity-driven flow systems. Epigenic unconfined karst is exclusively formed by gravity flow, but hypogenic speleogenesis is often a part of mixed flow systems, where groundwater flow is a result of different energy sources acting simultaneously, most commonly topography-driven flow and flow driven by temperature or solute density gradients. Flow driven by sediment compaction and tectonic compression can also contribute to mixed systems relevant to hypogenic speleogenesis, although the former is commonly volumetrically limited and the latter is temporally limited. A common misconception about hypogenic karst is that it is believed to be unrelated or contrasted to *meteoric* circulation (*e.g.* Budd, Saller, and Harris, 1995). Although non-meteoric waters (such as connate or magmatic waters) may be involved in some cases, most hypogenic speleogenesis is produced by predominantly meteoric waters, even where non-gravity drives for flow are involved. Meteoric waters in intermediate and deep (regional) flow systems, coming from distant recharge areas, can maintain or rejuvenate aggressiveness when entering a soluble formation from below in discharge areas to generate features that fall into the hypogenic class as it is defined above.

Segments of groundwater flow systems are characterized by three distinctly different flow regimes: the recharge, midline or throughflow, and discharge, with respective distinct physical, chemical, and hydrokinetic conditions. Hence, the rates and products (solutional porosity styles and patterns) of speleogenesis differ accordingly in karst systems associated with respective situations. In regions with substantial relief, composite flow patterns develop where flow systems at local, intermediate, and regional scales (types) are recognized. Figure 1, which is an adopted and modified version of Figure 1 of Tóth (1999), illustrates these flow regimes and types, with epigenic and hypogenic karst systems shown in the context of regional hydrogeology. In terms of regional hydrogeology, epigenic and hypogenic karst systems are regularly associated with different types and segments of flow systems. Epigenic karst systems are predominantly local flow systems, and/or parts of recharge segments of intermediate and regional flow systems. The recharge regime is characterized by relatively high hydraulic heads, decreasing with depth, and by downward and divergent flow. Hypogenic karst systems are associated with the discharge regimes of regional or intermediate flow systems, with largely the opposite energy and flow conditions. In discharge areas, hydraulic heads are relatively low and decrease upward, resulting in converging and ascending flow. However, in intermediate to regional confined systems, cross-formational communication causes recharge and discharge regimes (areas of correspondingly descending and ascending cross-formational communication) to laterally alternate in the throughflow area, largely following the gross topography (Shestopalov, 1981, 1989). For particular aquifers in multiple-aquifer systems, the relationships between recharge and discharge regimes are even more complex, with vertically superimposed recharge and discharge regimes (Section 3.1). From the perspective of a single formation or a bed, recharge includes all the ways that fluids enter the strata (Sharp and Kyle, 1988).

Recharge and discharge areas of basinal groundwater flow systems also have characteristic distinctions in groundwater chemistry and thermal regime. Groundwaters in recharge areas are typically chemically aggressive and promote dissolution, have low TDS, and are characterized by oxidizing conditions and negative anomalies of geothermal heat and gradient. In basinal groundwater flow systems there are systematic changes in hydrosomes with depth, from HCO_3 through SO_4 to Cl. Discharge areas have largely opposite characteristics: high TDS, chemical precipitation, accumulation of transported mineral matter, reducing conditions, and positive anomalies of geothermal heat and gradient (Tóth, 1999).

While recharge areas display highly variable input parameters in the groundwater regime (both hydraulic and chemical), in basinal discharge areas these parameters vary little over time and have low dependence on climate.

The general characteristics of the discharge regime, in geochemical terms, may not seem to favor speleogenesis. However, as demonstrated throughout this book, hypogenic speleogenesis is commonly associated not with terminal discharge regimes of basinal groundwater flow systems but with intermediate discharge limbs of these systems. Even more importantly, the fundamental feature of hypogenic speleogenesis is that it is driven by upward cross-formational hydraulic communication, so that relatively deep fluids of intermediate and regional groundwater flow systems interact with contrasting regimes of shallower aquifers and local systems. This causes disequilibrium conditions and favors various dissolutional mechanisms. These aspects are discussed elsewhere throughout this book.

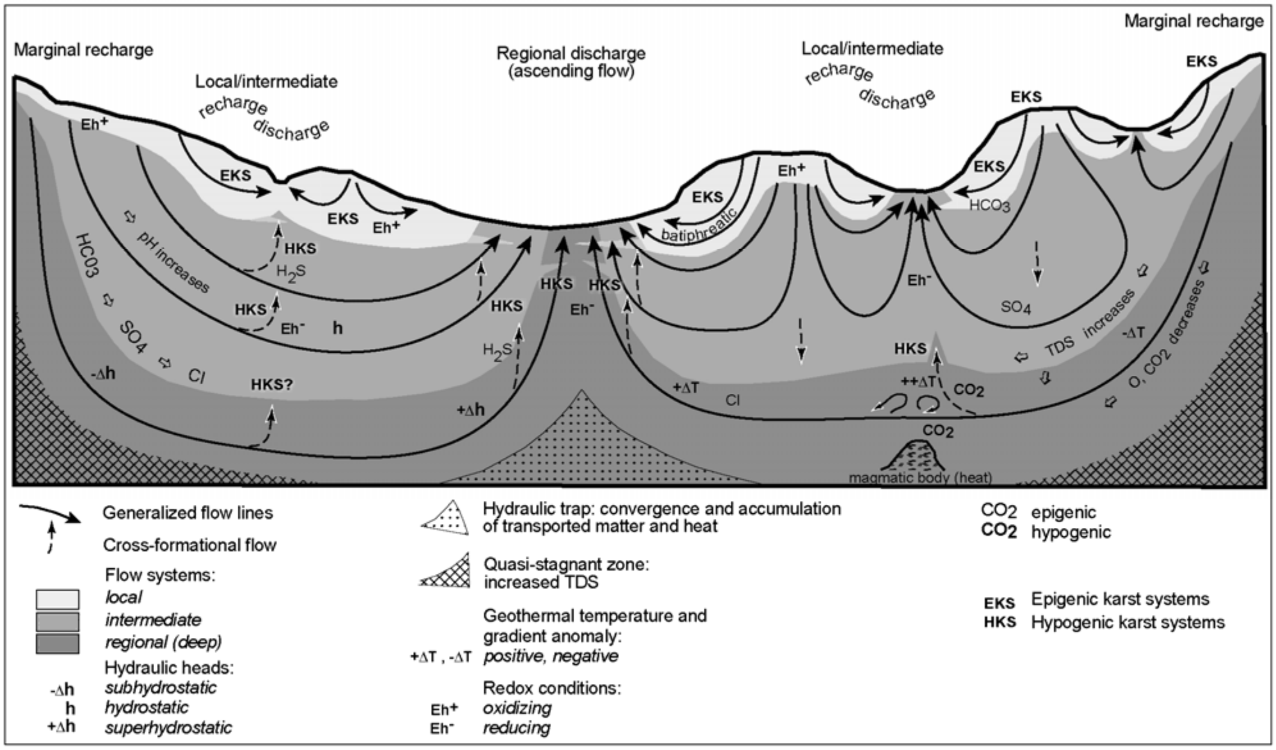

Figure 1. Epigenic and hypogenic karst in the context of basinal groundwater flow. Adopted and modified from Tóth (1999). The figure shows mainly gravity-driven flow in an idealized homogenous basin. In reality, most sedimentary sequences are highly heterogeneous, and gravity-driven flow interacts with other flow mechanisms.

As speleogenesis is a dynamic process capable of considerably changing primary porosity and permeability, it can itself create zones of high permeability along initially insignificant (in terms of regional or intermediate groundwater flow systems) cross-formational flow paths, or even without any initially guiding disruptions. In soluble beds (which originally commonly serve as confining beds; see Section 3.2) this goes through one of the mechanisms of ascending transverse speleogenesis, and in overlying insoluble beds permeability enhancement occurs via fracturing and brecciation above cave porosity zones. Thus, hypogenic speleogenesis may give rise to new discharge zones and contribute to segmenting laterally extensive "throughflow" regions.

The classification of regional hydrogeologic structures, introduced at the beginning of this section, can be presented as an evolutionary succession: artesian basins - disrupted basins - layered massifs - hydrogeologic massifs. This corresponds to the successive stages in the general tectonic and geomorphic evolution of sedimentary basins. Similarly, this evolutionary trend provides a framework to classify karst types and speleogenetic environments based on the evolutionary history of a soluble-rock aquifer (Klimchouk, 1996a; Klimchouk and Ford, 2000; Figure 2): from deposition and early emergence above sea level (*syngenetic /eogenetic karst*) through deep burial and re-emergence (the group of intrastratal karst types: *deep-seated karst, subjacent karst, entrenched karst*) to complete exposure (*denuded karst*). If karst bypasses burial, or if the soluble rock is exposed after burial without having experienced any significant karstification during burial, it represents the *open karst* type. Different types of karst, which concurrently represent the stages of karst development, are marked by distinct combinations of the structural prerequisites for groundwater flow and speleogenesis, flow regimes, recharge/discharge configurations, groundwater chemistry, and degree of inheritance from earlier conditions.

Although this classification does not directly specify the origin of caves, it characterizes dominant speleogenetic environments and their evolutionary changes. Karst types are viewed as stages of hydrogeologic/geomorphic evolution, between which the major boundary conditions, the overall circulation pattern, and extrinsic factors and intrinsic mechanisms of karst development appear to change considerably. The classification of karst types correlates well with the three major types of speleogenetic settings discussed above. Coastal and oceanic speleogenesis in diagenetically immature rocks falls into the syngenetic/eogenetic karst domain. Deep-seated karst is almost exclusively hypogenic. Subjacent karst can be induced by both, hypogenic and/or epigenic speleogenesis, depending on the scale of the flow system. Entrenched and denuded karst types are predominantly epigenic, with possible inheritance of hypogenic features, which are relicts in most of these cases. Open karst has exclusively epigenic speleogenesis.

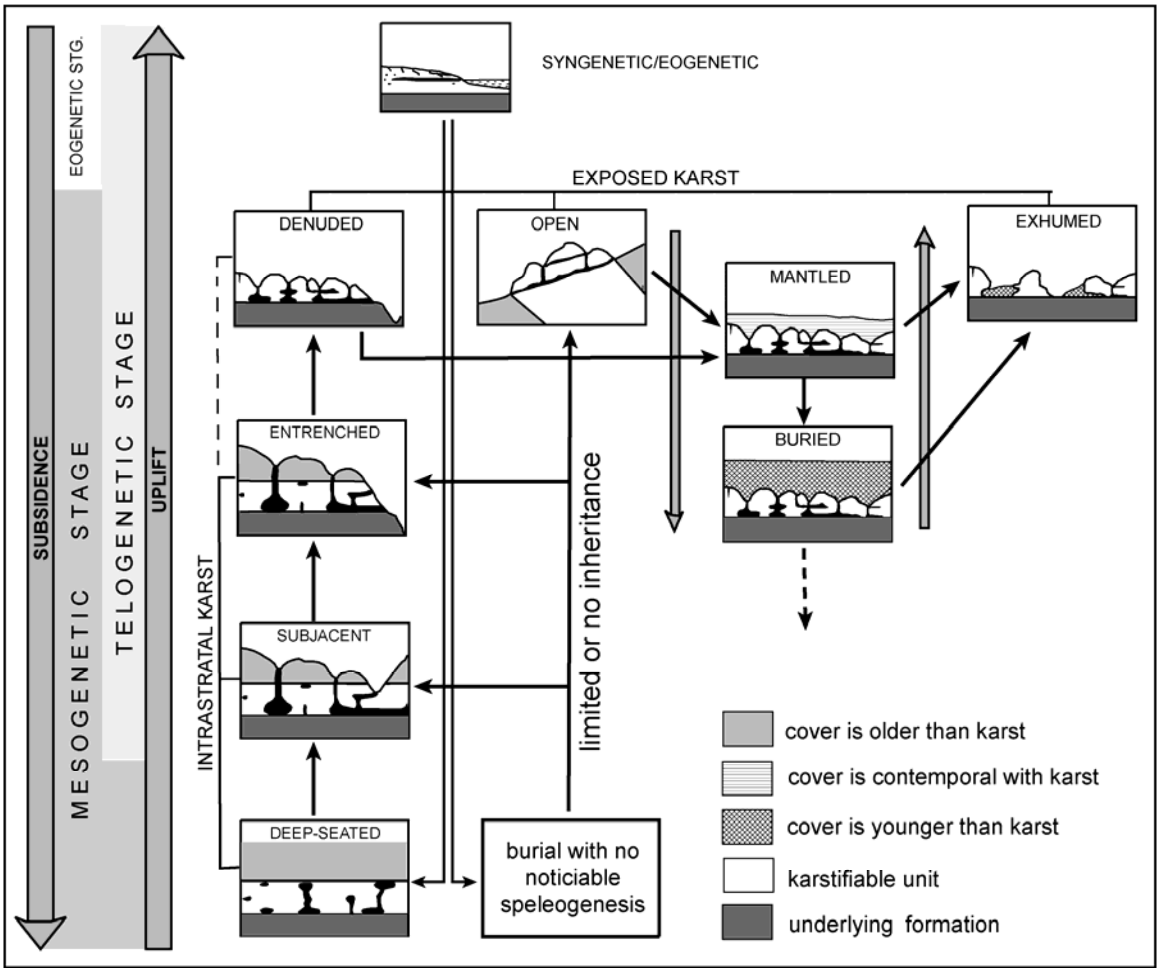

Figure 2. Evolutionary types of karst and speleogenetic environments (from Klimchouk and Ford, 2000).

3. Ascending hypogenic speleogenesis

In karst and cave science, three major problems can be traced that hindered proper understanding of hypogenic speleogenesis. First, caves formed in hypogenic and confined domains are accessible for exploration and study largely when they are brought into the unconfined realm due to uplift and denudation, hence when they become relict, decoupled from their formational environment and often partly overprinted by unconfined speleogenesis. They were commonly interpreted, despite now apparent contradictions, in the context of contemporary epigenic conditions. Classical examples are the works of Davis (1930) and Bretz (1942) who, in their theorizing of unconfined speleogenetic conditions, included many caves now known to have an ascending water origin (Ford, 2006). Although some of the most remarkable “atypical” caves are now recognized to be of hypogenic origin (*e.g.* Wind and Jewel caves of the Black Hills, South Dakota, USA; Carlsbad Cavern, Lechuguilla Cave, and other caves of the Guadalupe Mountains, USA; giant gypsum mazes of western Ukraine), many other caves await reevaluation. Another example is the former interpretation of the western Ukrainian giant mazes as being formed by lateral flow through the gypsum bed between entrenching sub-parallel river valleys (Dublyansky and Smol'nikov, 1969; Dublyansky and Lomaev, 1980). Second, even where a hypogenic origin was assumed, speleogenetic regularities and models devised for unconfined conditions were often simply taken to be equally applicable to the largely confined realm. And third, in earlier attempts to interpret some caves (particularly network mazes) in terms of artesian origin, old simplistic views of artesian flow were commonly implied, which again led to apparent unresolved contradictions.

Most stratified sedimentary basins are characterized by considerable heterogeneity and large contrasts in vertical permeability, which is, along with basin geometry conditions, the main cause of the wide occurrence of multiple-aquifer confined systems. The terms “confined” and “artesian” refer to hydrodynamic conditions and imply that groundwater is under pressure in a bed or stratum confined above and below by units of distinctly lower permeability. The potentiometric surface in such aquifers lies above the bottom of the upper confining bed. These terms are commonly used as synonymous and this usage is adopted here, although “artesian” was originally applied to aquifers in which the potentiometric surface lies above ground level.

Confusion often arises when the terms “artesian”, (“confined”) and “phreatic” are misleadingly understood as being equivalent, especially where bathyphreatic conditions are concerned. The term “phreatic” refers to conditions where water saturates all voids in a rock or sediment, in contrast to vadose conditions, above the water table, where voids are water-filled only temporarily, if ever. In this sense, phreatic unconfined and confined conditions are alike. Moreover, water in phreatic conduits is always confined by the host rock and possesses some hydraulic head above the conduit ceiling. 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.

Klimchouk (2000a; 2003a) suggested limiting use of the term “artesian” (“confined”) to prevailing flow conditions in an aquifer or a multiple-aquifer system, rather then to flow conditions within a single conduit. Use of the term “phreatic” should be restricted to the lower zone in unconfined aquifers, limited above by a water table that is free to rise and fall. The distinction between phreatic and confined conditions is important in the context of speleogenesis (see Section 3.7).

A new theory of speleogenesis in a confined multiple-aquifer system has been developed during the last two decades (Klimchouk, 1990; 1992; 1997a, 2000a; 2003a; 2004). It is based on:

1) Views about close cross-formation communication between aquifers in multi-story systems,

2) Ideas of hydrostratigraphic conversion of soluble formations in the system,

3) The concept of ascending transverse speleogenesis, and

4) Recognition of the ultimate control on confined speleogenesis by transmissivities of adjacent non-soluble aquifers.

3.1 Cross-formational communication and basinal hydraulic continuity

The older simplistic notion of artesian flow assumed that recharge to confined aquifers occurs only in limited areas where they crop out at the surface at higher elevations (*e.g.* at basin margins), and that groundwaters move laterally through separate aquifers within the throughflow area with no appreciable communication with adjacent aquifers across confining beds. Until recently, such views were commonly adopted in karst literature, resulting in one of the interpretative problems about artesian speleogenesis, mentioned above. This notion does not allow placement of artesian speleogenesis into the category of hypogenic speleogenesis as defined earlier. However, the most essential problem with it is that the implied substantial flow distances and travel times through soluble rocks generally preclude the possibility for significant conduit development in the confined flow area due to dissolution capacity constraints.

Since the middle of the 20th Century, close cross-formational communication between aquifers and basin-wide hydraulic continuity have been acknowledged in mainstream hydrogeology. It is well recognized that there are virtually no completely impervious rocks or sediments, just large contrasts in permeabilities. In modern hydrogeology the term “confined aquifer” is not used in the absolute sense of hydraulic isolation; a notion of semi-confinement is more appropriate as separating aquitards are commonly leaky at certain time and space scales.

Where there is a vertical head gradient between aquifers in a layered aquifer system, flow in high-permeability beds is predominantly lateral but flow in the separating low-permeability beds is predominantly vertical, if permeabilities differ by more than two orders of magnitude (Girinsky, 1947). Further developing these ideas, Mjatiev (1947) recognized that recharge areas of a confined aquifer are not just the uplifted marginal outcrops, but include all of the areas within the basin where the head is lower than in any adjacent aquifers. In the western literature, it was the work of Hantush and Jacob (1955) that introduced a “leakage factor” to account for hydraulic communication across confining strata and replaced the “confined aquifer” with the “multiple aquifer.” The concept of basin-wide hydraulic continuity has since become well accepted; the importance of cross-formational communication between aquifers has been recognized on a local scale from numerous aquifer and well data, and on a regional scale from basin hydraulics and water-resources evaluations. Shestopalov (1981, 1988; 1989) and Tóth (1995) provided important reviews and discussion of these characteristics.

This concept implies complex flow patterns in artesian basins and complex recharge-discharge characteristics for particular aquifers in the system. Besides marginal recharge areas and lateral flow components, this pattern includes laterally alternating recharge and discharge areas (areas of correspondingly descending and ascending cross communication) in the confined flow region, juxtaposition of recharge-discharge regimes for particular aquifers in a system, and flow systems at different scales.

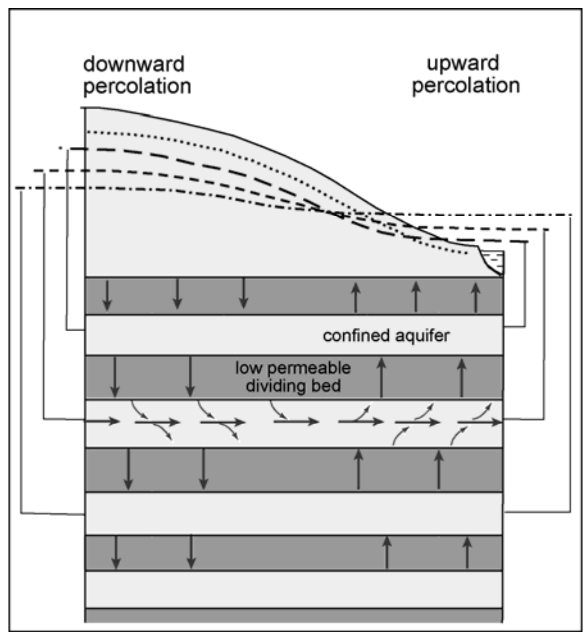

Figure 3. Flow pattern in a multi-story artesian aquifer system (from Shestopalov, 1989).

Figure 3 illustrates flow patterns in a typical multiple-aquifer system. Recharge to, and discharge from, a given aquifer may take place across intervening aquitards throughout the whole throughflow area. The amount and direction of hydraulic communication across homogenous dividing beds of low permeability depends on the relationships of heads between adjacent aquifers, which, in turn, are guided significantly by surface topography. For a given aquifer, there is a gradual vertical transition between net recharge and discharge, both of which occur simultaneously. At the regional scale, the respective areas of potentiometric highs and lows are the areas of downward and upward percolation, or zones of recharge and discharge. Potentiometric highs correspond to topographic highs, whereas potentiometric lows coincide with topographic lows, most commonly river valleys.

Detailed studies of regional hydraulics in many artesian basins around the world strongly support the above general views. A brief summary of basinal hydraulics that follows is derived from thorough regional studies of the major Ukrainian (Shestopalov, 1981, 1988, 1989) and Canadian (Tóth, 1963; Hitchon, 1969a, 1969b) basins to illustrate some important implications to hypogenic speleogenesis.

At the basin scale, the overall surface topography, along with lithologic and structural factors, controls distribution of potentiometric highs and lows and determines directions of cross-formational communication across confining beds. Downward percolation prevails along major divides, whereas notable upward flow is characteristic of areas below river valleys and other topographic lows. Regional confining formations of low permeability reduce cross-formation communication within multiple-aquifer systems, but do not isolate aquifers completely. The effects of topography can be traced to considerable depths in artesian basins. In particular, the draining influence of major rivers commonly extends to depths of 1000-1500 m and more (Shestopalov, 1989; Hitchon, 1969a; Figure 4).

Permeabilities of confining beds generally decrease with increasing depth. However, beneath river valleys the permeabilities in confining beds are considerably (up to 10 times) greater compared to adjacent areas, due to the fact that valleys often develop in tectonically weakened and disrupted zones, and/or there is relaxation of rocks from the load of formations removed by valley erosion. Ascending transverse speleogenesis in soluble formations, focused in the valley areas, and subsequent disruptions in formations that lie above soluble ones, may further contribute to this effect.

The rate of lateral flow through aquifers decreases with depth, from the margins toward the interiors of basins. However, beneath valleys a notable increase in flow rates is characteristic even for the central parts of the basins. Despite the fact that permeabilities of confining beds diminish with depth, the rate of vertical water exchange remains more stable than the rate of lateral flow, so that the relative significance of cross-formation flow increases with depth. The rate of upward percolation across confining beds under valleys is generally higher than the rate of downward percolation in the vicinity of topographic divides.

The influence of adjacent structural uplifts and massifs on basinal hydraulics is not only that the marginal recharge areas are located there, but that within the edges of basins there are particularly favorable conditions for enhanced cross-formational communication between aquifers. This is due to: 1) decreasing depths of aquifers and general reduction in thickness of sedimentary cover; 2) gradual transition of clay-rich facies to sandy-carbonate ones in confining beds; and 3) increase of fracturing due to weathering and unloading toward the basin edges (Shestopalov, 1989). Hitchon (1969b) discussed the effects of rapid recent uplifts in terms of upsetting the previously attained steady-state equilibrium of fluid flow in a basin.

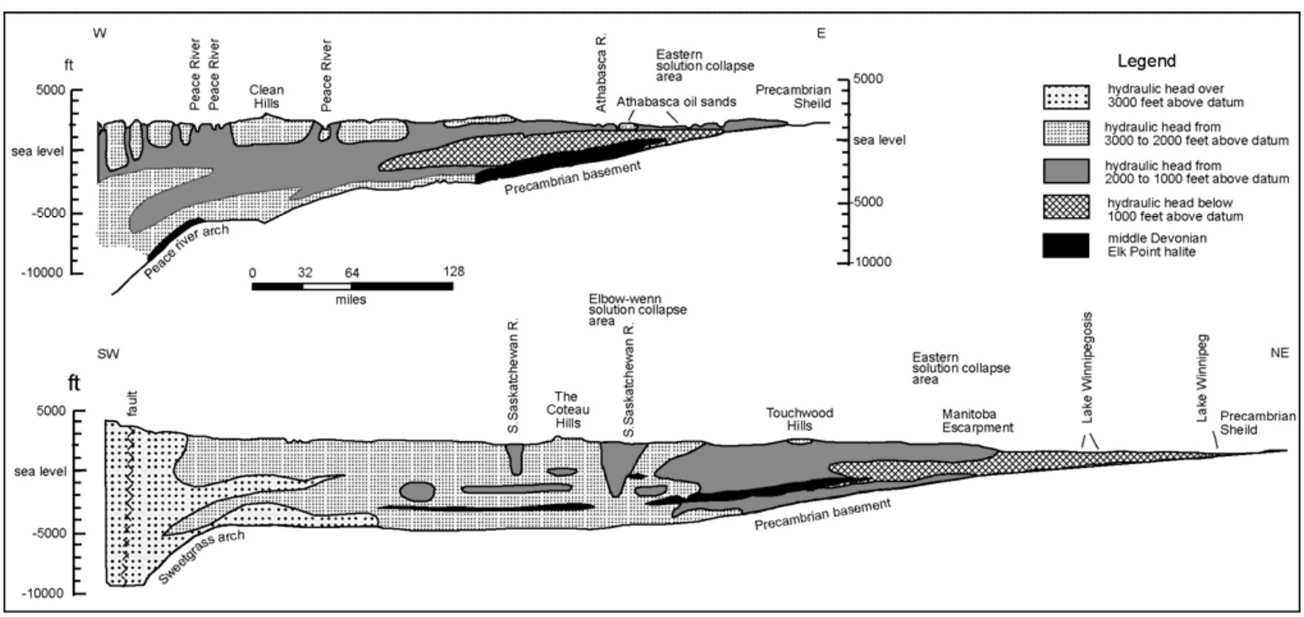

Figure 4. Hydraulic head distribution in cross sections, the Western Canada Sedimentary Basin showing the influence of topography, particularly river valleys (redrawn from Hitchon, 1969a).

The above general pattern is commonly complicated by faults, fault-cored anticlines, structural domes, stratigraphic "windows," karstified zones, etc. These features create additional preferential paths for focused cross-formational communication, which sometimes considerably complicate the regional hydrogeologic pattern. The overall importance of cross-formational flow can be illustrated by the following estimates for the Dnieper-Donetsk basin (eastern Ukraine), where the flow in aquifers is supported largely by vertical water exchange (up to 88-100%) rather than by lateral communication with marginal recharge areas. In the central parts of the basin, lateral inflow from the adjacent areas comprises only 10 to 32% of the total groundwater flux (Shestopalov, 1989).

Hitchon provided an excellent study of basin-wide distribution of hydraulic heads in the Western Canada Sedimentary Basin, particularly illustrative for both topographic (1969a) and geologic (1969b) effects. Both major and minor topographic features are shown to exert an important control on the distribution of recharge and discharge areas and on the regional and local flow systems. However, the variations of geology, particularly the presence of highly permeable beds, even quite local laterally, result in significant changes in the regional topography-induced flow system. It was demonstrated in Hitchon (1969b) that a pattern of low fluid potential areas is largely related to distribution of the Upper Devonian and Carboniferous carbonate and evaporite rocks, particularly of carbonate reef complexes. However, the fluid potential lows and highs are recognized within the carbonate rocks themselves, indicating laterally uneven permeability distributions. Even lithofacies changes within a major soluble formation, such as variations in proportions of anhydrite in carbonate rocks or changes from dolomite to limestone, are often reflected at the basinal scale in the hydraulic head maps for a given formation and juxtaposed sequences. Some reef complexes are known to lack good, high permeability continuity in an updip direction but have excellent hydraulic continuity between the vertically adjacent reef complexes. Drawdown imposed in some of the low-fluid-potential areas is reflected through up to 900 m of strata, across several stratigraphic units and a major unconformity. Clearly, many geology-induced variations of hydraulic head distribution within the basin, as described by Hitchon (1969b), could be best interpreted in terms of, and shown to illustrate the role in the basinal hydrogeology, of hypogenic transverse speleogenesis as it is treated in this book. See Figure 5 for a case from Alberta, which can be conceptually translated to the buried Capitan reef complex in the eastern Delaware Basin of the USA, with associated transverse karstification in both the reef and the overlying evaporite strata (Figure 59). The model speleogenetic reference for both situations could be the known hypogenic caves in the presently exhumed part of the Capitan reef complex in the Guadalupe Mountains.

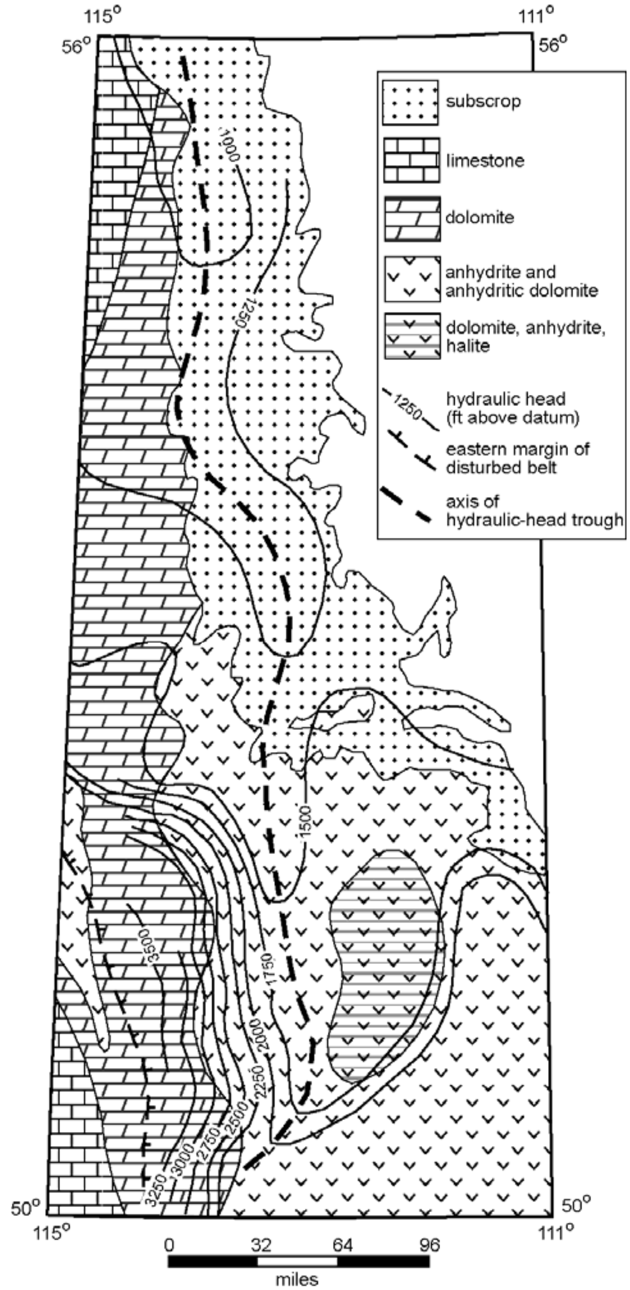

Figure 5. Hydraulic head distribution in the Upper Devonian Wabamun Group of central and southern Alberta (from Hitchon, 1969b). As described in that paper, the north-trending trough on the potentiometric surface in the subscrop of this group, shown on the figure, is reflected in a corresponding trough on the potentiometric surface of the overlying Manville Group.
A corresponding trough can be seen in the underlying Winterburn Group, which in turn overlies the low fluid potential drain formed by the Crossmont reef complex, a hydraulic extension of the low fluid potential system of the Rimbey-Meadowbrook reef chain.

Another important generalization from basin-wide studies is that rates of cross-formational exchange in a multiple-aquifer system depend not only on permeabilities, thicknesses, continuity, and number of intervening confining beds, but also on the tectonic regime of a region. The uplift trend and the neotectonic activity favor cross-formational communications between aquifers.

Cross-formational hydraulic communication is one of the most important factors determining the resources and chemical composition of groundwaters in the upper hydrogeodynamic stories of sedimentary basins. It has been largely overlooked in karst hydrogeology and speleogenetic studies. It obviously has an immense importance, and provides a broad perspective for implications to hypogenic speleogenesis. The above described regularities of basinal hydraulics determine locations of zones particularly favorable for speleogenesis.

The place of hypogenic speleogenesis within a basinal flow domain is shown in Figure 1. It is commonly associated with discharge segments of regional or intermediate flow systems. But arguments throughout this book, supported by numerous field examples worldwide, strongly suggest that this association is largely because hypogenic transverse speleogenesis creates these discharge segments, making them recognizable at the regional scale, by greatly enhancing initially available cross-formational permeability structures or even by creating new efficient ones. The profound effect of a high permeability rock unit on the regional flow pattern was demonstrated theoretically by Freeze and Witherspoon (1967) and validated by many regional hydrogeologic studies. In basins containing soluble formations, the primary result of speleogenesis is converging groundwater flow to zones where ascending cross-formational communication is greatly enhanced by speleogenesis.

3.2 Hydrostratigraphic conversion of soluble formations

The hydrostratigraphy of a sedimentary succession is determined mainly by the relative permeabilities of the rock units. Aquifers are separated from each other and from the upper unconfined aquifer by low-permeability beds. Initial matrix permeabilities of common porous aquifers (*e.g.* many medium- to coarse-grained clastic sediments) are normally several orders of magnitude greater then that of soluble rocks such as massive limestones or sulfates prior to speleogenesis. Soluble formations are commonly vertically conterminous with formations with initially higher permeability and serve as intervening beds (aquitards) in a multiple-aquifer system. However, they change their hydrostratigraphic role to karstic aquifers in the course of speleogenetic evolution.

As tectonism and uplift impose fracture permeability, intervening soluble units increasingly transmit groundwater between (non-karstic) aquifers in zones of sufficient head gradients. According to Girinsky's (1947) premise, flow in separating beds is predominantly vertical, so speleogenesis evolves in transverse communication paths across the soluble formation. When conduit systems are developed within soluble formations, conventional karst wisdom views the situation as a karst aquifer sandwiched between relative aquitards, without appreciating that the initial conditions were quite the opposite (Figure 6). Failure to recognize the proper relationships commonly causes reverse misconceptions in hydrostratigraphic interpretations of layered systems: soluble units, particularly evaporites, may be treated as impervious beds (aquitards) in a system, or they may be regarded "by definition" as karstified media.

Switching of hydrogeological functions of different beds in a sequence during the speleogenetic evolution of the soluble ones is quite common in confined settings (Klimchouk, 1990, 1992, 1994, 1996b, 2000a; Lowe, 1992). This is because changes in permeability of soluble units through time are much more dynamic and drastic than changes in non-soluble beds. However, it is important to recognize that hydraulic properties of karst aquifers evolved in response to hypogenic transverse speleogenesis, and are characteristically different from epigenic karst aquifers, the aspects further discussed in the next three sections.

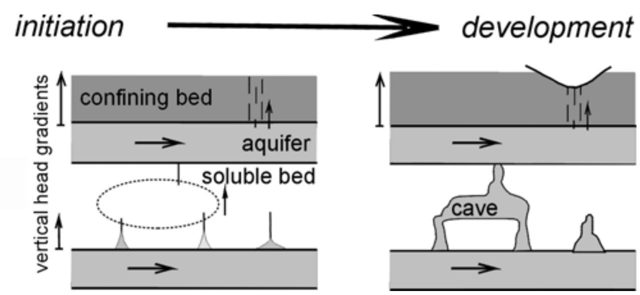

Figure 6. Conversion of the hydrogeological function of a soluble bed in a multiple-aquifer system in the course of speleogenesis (from Klimchouk, 1996b).

3.3 The concept of transverse speleogenesis

In unconfined settings, early conduit development occurs by lateral flow through an aquifer, from the input boundary to the output boundary. Furthermore, it was commonly implied that water flows along the long dimension of fractures, which are commonly arranged laterally relative to bedding (Figure 7), or along pathways that combine long dimensions of several laterally connected fissures. Long flow lengths and therefore low discharge/length ratios (*sensu* Palmer, 1991) are inferred in such a configuration, which is commonly used in modeling of early conduit development. Similarly, the parameter of passage length, or cave development, derived from speleological mapping, tacitly implies the meaning of the length of flow that formed a passage. These views represent what can be called *lateral (or longitudinal) speleogenesis*, a concept that is generally adequate when applied to unconfined settings. It is deeply rooted in the speleogenetic literature and was commonly translated to speleogenesis in confined settings within the older simplistic artesian flow concept, resulting in misleading implications.

As shown above, ascending hydraulic communication across compact soluble beds is predominant in leaky confined multiple-aquifer systems. However, the conventional concept of lateral speleogenesis does not adequately reflect the arrangement of flowpaths and the flow pattern in this case. Klimchouk (2000a; 2003a) suggested a concept of *transverse speleogenesis* to describe ascending conduit development in a soluble formation recharged from below.

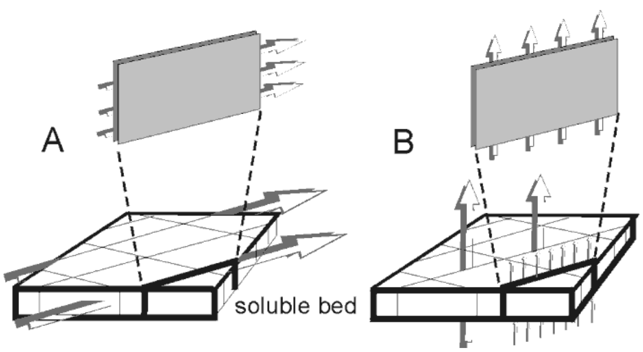

Figure 7. A diagram illustrating general concepts of lateral (A) versus ascending transverse (B) flow through a single fracture and a fissure network encased in a soluble bed (from Klimchouk, 2003a). See also Figure 8.

Where upward flow occurs through a fractured soluble bed that functions as a leaky aquitard, flow actually follows the fracture height (Figures 7 and 8-A), or along a sequence of heights of the vertically connected fractures (Figure 8-B). Flow distances through a soluble rock are rather short, commonly on the order of meters or a few tens of meters, thus allowing rather high discharge/length ratios. Where laterally continuous fracture networks are present within certain intervals, flow and speleogenesis may include a lateral component within the generally transverse flow pattern. Maps of caves formed in this way may display tens or even a few hundred kilometers of integrated passages, spread over hundreds of meters of straight lateral distance, but these figures have nothing to do with the actual flow pattern and flow length through the soluble formation in the case of transverse speleogenesis in confined settings.

Transverse speleogenesis denotes conduit development driven by the vertical gradients in hydraulic head or density across a soluble formation so that flow is generally directed transversely relative to bedding and stratiform fracture networks, often arranged in several superimposed stories (Figure 8-B). The pattern of transverse speleogenesis may include some lateral components within laterally extensive and connected permeability structures, but the overall cave-forming flow system remains transverse relative to the soluble formation, linking together its vertically arranged input and output boundaries.

It is interesting to note that speleogenesis in the vadose zone is also predominantly transverse in the sense described above. However, the main regularities and patterns of speleogenesis are strikingly different in the case of descending free-flow through the vadose zone than in the case of ascending flow through a confined system.

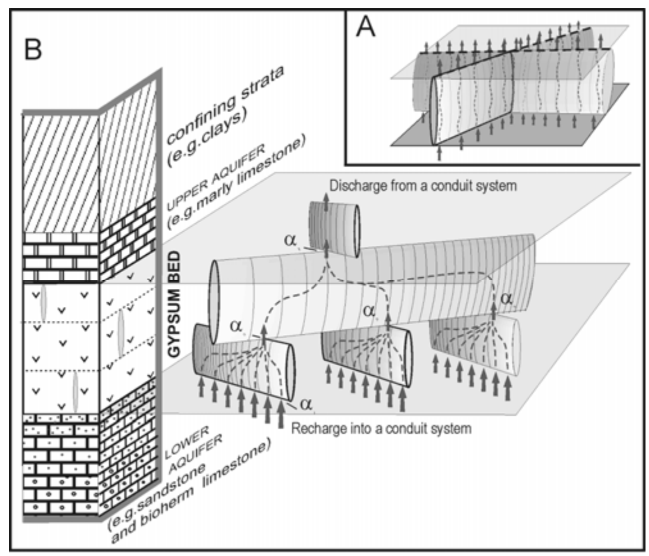

Figure 8. A = Transverse flow through a fracture network in a single level, with fractures crossing a bed for the whole thickness; B = transverse flow through fracture networks in multiple levels. Litho- and hydrostratigraphy depicted corresponds to the case of the western Ukraine, however such multi-level arrangement of fracture networks is common for stratified carbonate and sulfate sequences (from Klimchouk, 2003a).

In describing the concept of ascending transverse speleogenesis, reference to bedding implies "idealized" settings of sub-horizontal stratification predominant in most basins. However, it can be misleading in the case of intensely folded formations, sub-vertical bedding common in folded regions, and ancient structural stories of many basins that experienced a long and complex evolution. See Osborne (2001a) for a case of ascending speleogenesis in steeply-dipping rocks. In such situations, the concept should be better taken with reference to sequence boundaries or simply to the approximate horizontal datum.

3.4 Vertical heterogeneity in porosity and permeability

Non-soluble rocks, as well as soluble rocks before speleogenesis commences, are commonly characterized by matrix and fracture porosity. In different lithologies and lithofacies, the relative significance of the porosity types, as well as permeabilities of respective media and interaction of respective flow systems, varies broadly depending on sediment type and diagenetic, tectonic and geomorphic history. Evaporites and mature carbonates normally have low matrix permeability and most flow is transmitted through fractures, with still greater concentration of flow in conduits when they evolve. Young carbonates have more diverse and generally greater matrix porosity, variously combined with fracture and conduit systems. The touching-vug porosity of Vacher and Mylroie (2002) is a specific sub-type of matrix porosity, with greatly enlarged and interconnected pores-vugs, or it can be considered as a sub-type of conduit porosity.

In a stratified system, vertical (layered) heterogeneity in the original porosity and permeability structure is commonly much greater than lateral. In addition to large contrasts in bulk permeabilities of adjacent beds and horizons, there are effects of limited connections between different juxtaposed stratiform porosity systems (Figure 8 and 10). All these vertical heterogeneities play an important role in configuring groundwater flow in general, and particularly in ascending transverse speleogenesis.

In the simplest case, there is a thin homogeneous fractured bed of limestone or gypsum, sandwiched between diffuse permeability aquifers, in which each fracture directly connects the bottom and top boundaries (Figure 8-A). This type of “sandwich aquifer,” where a thin carbonate unit is overlain and underlain by insoluble strata, has been described by White (1969), who noted that network caves are characteristic for this situation. In fact, actual patterns of resultant caves are strongly dependent on fracture distribution and arrangement. Network caves are formed where there is a continuous fracture network encased in the bed. If the soluble bed is only occasionally fractured, single, laterally-isolated fissure-like passages may form with both ends blind-terminated, or small clusters of several intersecting passages. Illustrative examples are caves encountered by mines in a thin Miocene limestone bed in the Prichernomorsky artesian basin, south Ukraine (see Figure 31).

Apart from major sedimentological heterogeneities in the vertical section, such as alternating prominent beds of contrasting lithologies that determine the principal hydrostratigraphy in a basin, depositional environments and facies changes within an otherwise “homogeneous” soluble formation also play an important role in determining secondary porosity and permeability distribution and their subsequent evolution through burial diagenesis and tectonism. Individual beds or formations commonly differ in nature, patterns, and frequency of fracture networks. Hence, these conditions will impose strong control on the structure of subsequent hypogenic speleogenesis.

In the Miocene gypsum formation in the western Ukraine, which hosts the giant artesian maze caves, the section is typically composed of two or three varieties of gypsum differing in texture and structure. Each bed encases laterally continuous extensive stratiform fracture networks, largely independent of the network encased by adjacent beds (Figure 8-B). Fracture orientation and frequency differ between the beds (Klimchouk *et al.*, 1995), so fractures in one bed are rarely co-planar with fractures in an adjacent bed, but they may have occasional vertical connections at discrete points. Such discordance in permeability structure between adjacent beds creates the flow constraint effect and causes some lateral component in the generally transverse flow. The same effect is caused by discordance in permeability structure and overall values between the lower and upper aquifers and respective adjacent beds in the gypsum bed. Because of the lateral component and good fracture connectivity at certain levels, integrated systems of passages develop on such master levels, which gives a misleading impression of generally lateral cave-forming flow through a soluble unit or its particular bed. Multi-story (three-dimensional) maze caves with stratiform levels formed in this way may have a few kilometers to a few hundreds of kilometers of laterally integrated passages, which further favors the misleading interpretation of the cave-forming flow to be lateral. In cases where laterally connected fracture networks are sub-horizontal, the resultant cave levels are commonly misinterpreted as levels in the evolutionary sense within the epigenic paradigm, *i.e.* abandoned tiers of phreatic development or cave levels at the water table. Another common misinterpretation of such levels is that the downward cave development was perched on the underlying non-soluble bed (which is now a low-permeability bed as compared to the already karstified soluble unit, but that used to be a feeding aquifer at the time of early speleogenesis – see notes on the conversion of the hydrostratigraphy above).

The above-described arrangement of the original (pre-speleogenetic) porosity is shown to be one of the main controls for transverse ascending speleogenesis and the structure of two to three story cave systems in the western Ukraine (Klimchouk and Rogozhnikov, 1982; Klimchouk, 1990 and 1992; Klimchouk *et al.*, 1995; see Figure 12). The structure of the multi-story mazes of Wind and Jewel caves in the Black Hills, South Dakota, USA is controlled largely in the same way (Ford, 1989), although bedding, superimposed stratiform fracture networks, and the resultant cave “levels” here are dipping.

The lithostratigraphic cyclicity of various scales is common within thick carbonate sedimentary successions of different ages. A number of studies demonstrate that the distribution of porosity and permeability relates directly to lithofacies, so that cyclostratigraphy is increasingly used for characterization of vertical heterogeneity of porosity and permeability (*e.g.* Hovorka *et al.*, 1998; Budd and Vacher, 2004).

Budd and Vacher (2004) show that matrix permeability of young carbonates in the Upper Floridan Aquifer ranges over three orders of magnitude between different lithofacies. Cunningham *et al.* (2006) developed a high-resolution cyclostratigraphic model for the carbonate Biscayne Aquifer, Florida, and demonstrated pronounced regular variations in porosity structure and permeability between lithofacies, arranged in cyclic successions of three types. Permeability of the aquifer is heterogeneous, with values differing up to two orders of magnitude between the lithofacies. Three types of flow zones, interbedded with low-permeability zones, are recognized. High permeability zones of "touching-vug" type (Vacher and Mylroie, 2002) provide stratiform passageways for groundwater flow. In the context of transverse speleogenesis, such diffusely permeable beds play the role of "aquifers" in heterogeneous sequences, with vertical flow and conduit development occurring in the intervening less permeable beds. This case of young eogenic carbonates is referred to in order to illustrate the importance of detailed hydrostratigraphic consideration for karst aquifers, which is equally important for older successions of well-indurated carbonate rocks. Unfortunately, such detailed hydrostratigraphic characterizations are not commonly available for multiple-aquifer systems containing soluble units.

In the Ordovician Knox carbonates in Tennessee, USA, the pattern of transgressive and regressive cycles creates pronounced hydrostratigraphic heterogeneity (Montanez, 1997). The facies within regressive cycles are almost completely replaced by tight, fine-crystalline dolomite that formed syndepositionally from evaporating water. They became aquitards afterwards. In contrast, transgressive cycles, which behaved like aquifers during burial diagenesis, have considerable porosity and permeability as they were affected by extensive dissolution in intermediate to deep-burial settings, according to petrographic data.

Machel (1999) refers to reef complex carbonates as another example for depositional control on diagenesis that leads to a pronounced differentiation into "proto-aquifers" and "proto-aquitards" at the time of deposition. In addition to the effect on diagenetic changes of porosity in different facies, subsequent fracturing develops in distinct styles and with different frequency in the facies varieties to further accentuate permeability differences. As a result, ascending transverse speleogenesis across such vertically heterogeneous sequences will utilize various kinds of original (pre-speleogenetic) porosity at different intervals (cross-bedding and stratiform fracturing of different styles, touching-vug porosity, etc.) and will be affected by constraints of their poor connectivity with porosity at other intervals. Complex 3-D structural organization of the ascending hypogenic caves in the Permian reef complex of the Guadalupe Mountains in the southwestern USA perfectly illustrates the effect of this heterogeneity (see Section 4.5). A small-scale illustration is presented by the photo of Figure 9 taken in Caverns of Sonora, Texas, USA. Numerous cupolas at the ceiling of the uppermost story of the multi-story cave (view from below) open up into a distinct horizon of touching-vugs type porosity ("burrowed bed"), which served as an outlet boundary for the uprising cave-forming cross-formational flow.

In many evaporitic sequences, gypsum beds of moderate thickness alternate with densely fractured limestone or dolomite beds that originally played a role of aquifers and lateral "carrier" beds. They supply water, undersaturated with respect to gypsum, to sub-gypsum beds, from where the water enters the gypsum, with upward flow driven by the hydraulic head gradient and/or density gradient (Stafford et al., 2008). With the onset of transverse speleogenesis, various units in the sequence receive better hydraulic connection and the whole sequence then behaves largely as a single multiple-aquifer system. This is the situation common for many mixed sulfate-carbonate sequences, *e.g.* the Permian sequences of the fore-Ural, Pinega, and North Dvina regions in Russia; and the Rustler and Seven Rivers Formations in the Delaware and Roswell basins in New Mexico, USA.

Recognition of the nature of transverse speleogenesis and aspects discussed above will help to develop more adequate approaches to flow models in stratified heterogeneous karst aquifers. Lateral transmission of groundwater occurs mainly through non-karstic aquifers, or through specific horizons of the laterally connected fracture networks or "touching-vug" type porosity within the soluble sequence. Ascending transverse cave development provides for vertical communication between such laterally conductive beds. It is important to recognize that because speleogenesis in layered confined systems evolves in response to generally transverse flow across soluble dividing (originally less-permeable) beds, the resultant conduit systems, even when mature, do not provide for significant lateral hydraulic connection at the basin scale. Even the largest maze systems in soluble beds have continuous lateral extent through areas of only a few square kilometers at a maximum, and for a few hundred meters in any single direction. In the lateral aspect they remain isolated clusters rather than laterally extensive systems. However, mature transverse conduit systems provide ideal local vertical hydraulic connections between lateral stratiform passageways for groundwater flow, and hence significantly affect basinal flow pattern.

Figure 9. A cupola at the ceiling of the uppermost story of Caverns of Sonora, Texas (view from below; breadth of the photo is about 2 m). Numerous cupolas at this story open up into a distinct horizon of touching-vugs type porosity (“burrowed bed”), which served as a receiving aquifer during the formation of the cave. Developed at four stories within a vertical range of about 35 m in the layered carbonate Edwards Group, the cave passages are mainly controlled by fractures encased in distinct beds of compact limestone, although some more prominent fractures, and hence passages, cross through several beds.

3.5 Recharge, cave-forming flow and discharge in hypogene settings

The mode of recharge and discharge, and relationships between the respective boundaries and zones, are among the major factors that determine the style of speleogenesis and resultant cave patterns. In hypogenic speleogenesis, recharge of water to a given soluble formation occurs from the adjacent formation below, the main criterion for distinguishing this class of speleogenesis. Another important difference is that discharge occurs also through non-soluble formations. Hydraulic properties of adjacent formations, and of a major upper confining formation, impose important controls on speleogenesis in confined settings (Klimchouk, 2000a; 2003a).

To prevent confusion that arises when referring to stratigraphic units in terms of their solubility, the formation that receives recharge from below and hosts hypogenic caves will be referred to as a *cave formation*, or a *cave unit*, the underlying source formation is a *feeding formation*, and the overlying formation into which discharge occurs is the *receiving formation*. All of the formations can be generally soluble, but still with drastically different capacities to support cave development under given physical, chemical and hydrokinetic conditions.

The mode of recharge, in terms of its lateral distribution, depends on the types, distribution and connectivity of the original porosity systems in both the feeding formation and the cave formation as well as the overall hydrostratigraphy of the system. In the feeding formation, effective porosity at the contact can be diffuse and homogenous (A1-A3 in Figure 11), diffuse and inhomogeneous (zones of enhanced conductivity in otherwise permeable media; B1-B3), or localized (tectonically disrupted zones, *e.g.* fault zones, etc.) In the latter case, high conductivity zones in the feeding and receiving formations are commonly co-planar with the respective major permeability paths across the whole cave formation (A4). More commonly, there is a disparity of permeability structures between the feeding and receiving formations.

Such a disparity is almost always the case at the lower contact of the cave formation, at its recharge boundary (Figure 10). Permeability in the feeding formation can be represented by various combinations of matrix, touching-vug, fracture or conduit (prominent cross-bedding fractures) porosity systems, but it never matches the original permeability structure in the cave formation. The extreme case of such a disparity is where there are virtually no hydraulically efficient original flowpaths available in the cave unit at its lower contact, and hence no perceptible forced flow through it (Figure 10, F-G). Still, hypogenic speleogenesis can operate in this situation through the free convection mechanism.

Vertically across the cave formation, original flow paths are almost always composed of segments of different porosity styles and types, as discussed and exemplified in the previous section. In fact, the relations discussed above and shown in Figure 10 may be found between adjacent horizons/units within the inhomogeneous cave formation itself (see Figure 9). In addition to distinct characteristics of different porosity segments, the connectivity constraints between them impose strong effects on speleogenesis. Complex 3-D structural organization of the most widely acknowledged ascending hypogenic caves, such as gypsum mazes in the western Ukraine, limestone mazes of the Black Hills, South Dakota, USA, composite pattern caves of the Buda Hills in Hungary and the Capitan reef complex of the Guadalupe Mountains, New Mexico, USA, illustrate the effect of this heterogeneity. It is important to underscore that because hypogenic speleogenesis is often the product of mixed (topography-driven and density-driven) flow systems, buoyancy effects are commonly involved in establishing hydraulic connections between different porosity segments (Section 3.8).

Discharge from the cave formation is commonly mediated by some kind of aquifer with diffuse permeability ("receiving formation"; rows 1 and 2 in Figure 11), but may also occur via localized zones of enhanced permeability in the immediately overlying leaky aquitard (rows 3 and 4). In gravity-driven flow systems, zones of upward flow establish themselves when hydraulic gradients across the separating beds, particularly across the major confining formation, are maximized below river valleys or other prominent topographic lows, and/or where zones of enhanced vertical permeability across the leaky aquitard facilitate discharge. The overall discharge from a system can be diffuse (A1 in Figure 11), or localized in single (B1, A3, B3, A4) or multiple (A2, B2) fault zones. Zones of preferential recharge to the cave formation and zones of overall discharge can be laterally shifted relative to each other, resulting in a staircase effect in the arrangement of a multi-story cave system, with offset of upper stories toward the focuses of discharge. This is exemplified by Jewel and Wind caves in the Black Hills (Ford, 1989) and by the Optymistychna caves (Klimchouk *et al.*, 1995), as shown in Figures 12 and 13.

For cross-formational flow the least permeable units dominate the system. Initially poorly permeable soluble beds may develop dramatically increased permeability due to transverse speleogenesis, although non-soluble beds do not. Figure 11 (see also Figure 6) illustrates an important feature of hypogenic, confined transverse speleogenesis, distinct from epigenic settings; when conduits have evolved (*i.e.* after kinetic breakthrough), the flow across the cave formation is limited by the permeability of the feeding and receiving formations and boundary conditions of the respective intermediate or regional flow system. This has important speleogenetic consequences, as it suppresses speleogenetic competition in the developing transverse system and favors the formation of pervasive cave patterns where proper structural prerequisites exist (Klimchouk, 2000a; 2003a; Birk *et al.*, 2003; see Section 3.7).

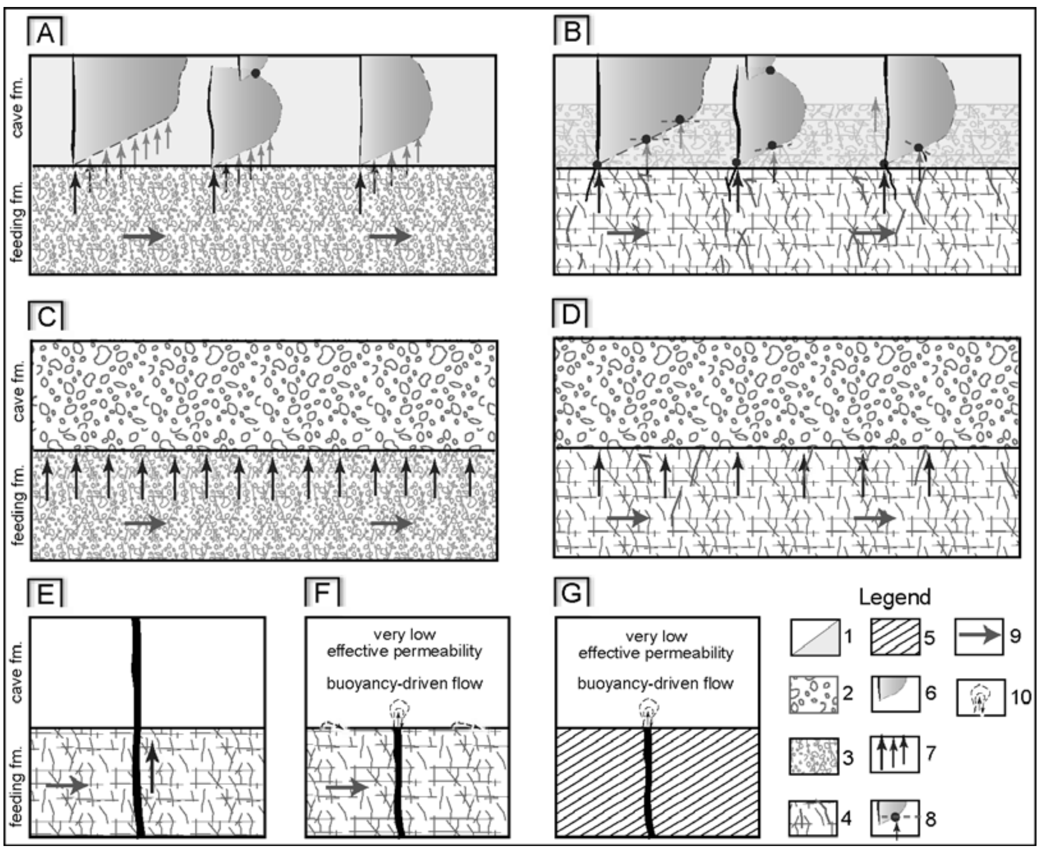

Figure 10. Modes of recharge to a cave formation from a feeding formation, depending on juxtaposed permeability structures. Note that similar relationships may occur between different beds in the cave formation, leading to complex 3-dimensional organization of ascending cave systems. Key to legend: permeability styles: 1 = soluble rocks of low matrix permeability; 2 = poorly connected vug-type porosity – low effective permeability; 3 = high matrix-vug permeability; 4 = high fracture permeability; 5 = insoluble rocks of low permeability; 6 = prominent fractures and their planes; 7 = recharge to a cave formation; 8 = points of fracture intersections; 9 = lateral flow through aquifers; 10 = density-driven dissolution.

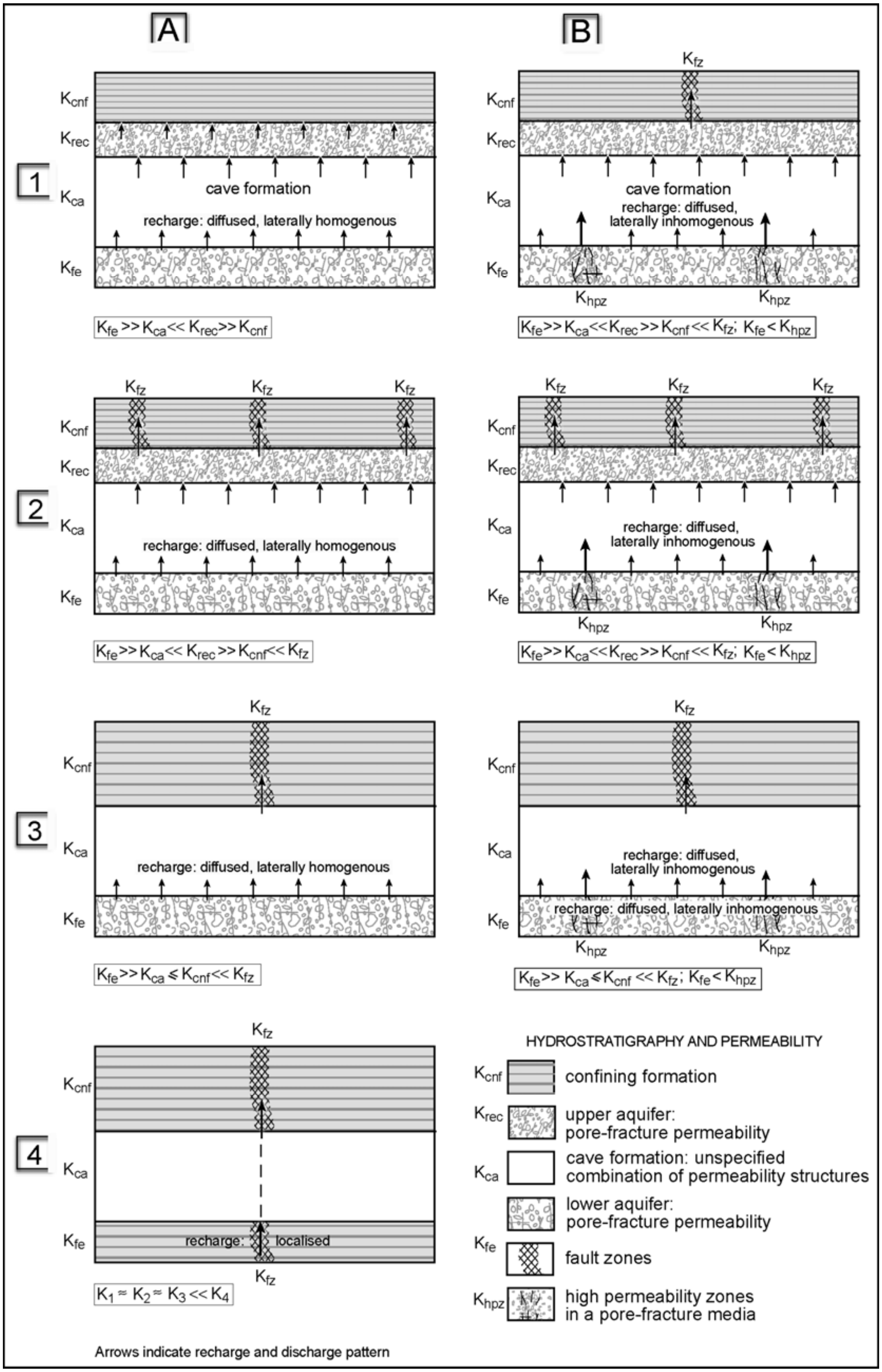

Figure 11. Variants of hydrostratigraphic position and recharge/discharge relationships for a cave formation in a layered aquifer system. The actual mode of recharge and discharge depends also on the permeability structure within the cave formation, which is not specified. It is assumed to be diffuse in all diagrams but A4. See text for further discussion.

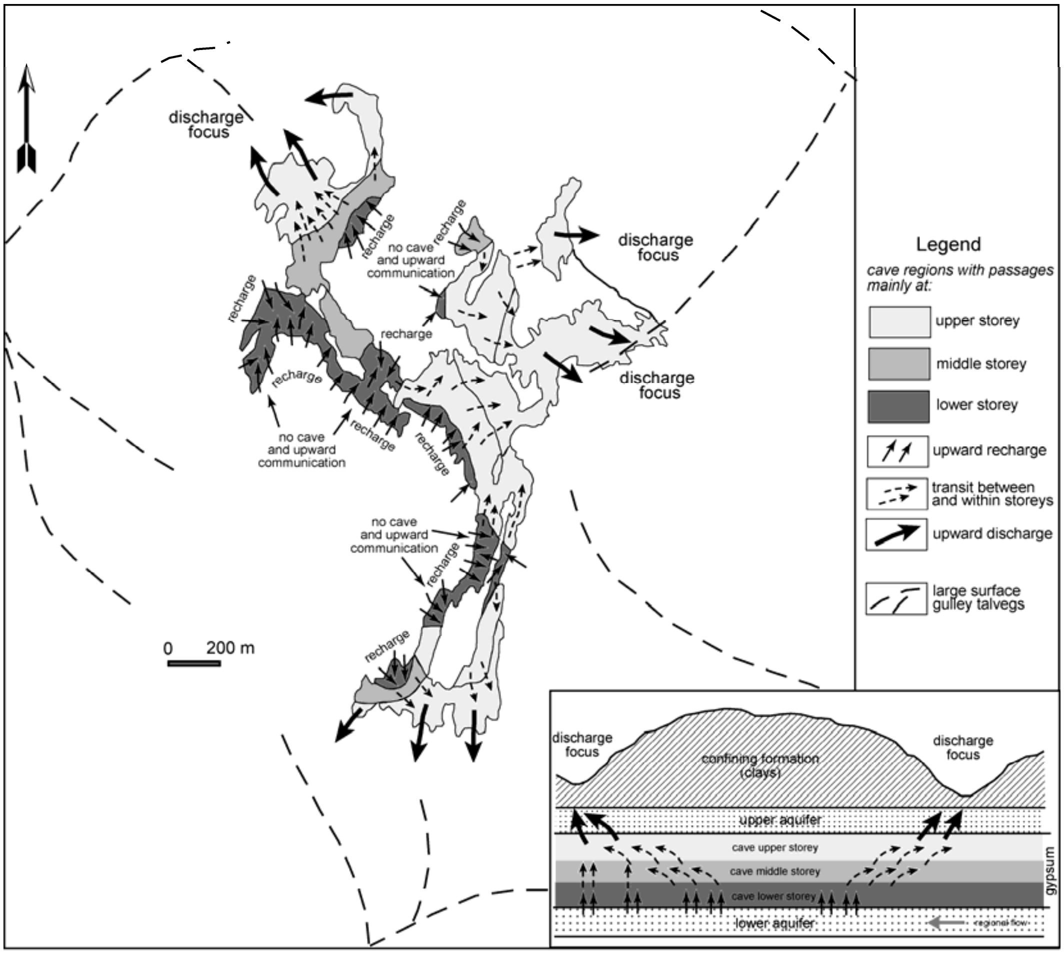

Figure 12. Regionalization of Optymistychna Cave, western Ukraine, a 214+ km long maze, according to its multi-story structures and the reconstruction of recharge-discharge arrangement during the time of its formation. Inset diagrammatically shows the hydrostratigraphy and the stair-case effect of the recharge-discharge offset. The thickness of the gypsum bed (18 m) is exaggerated. From Klimchouk *et al.* (1995).

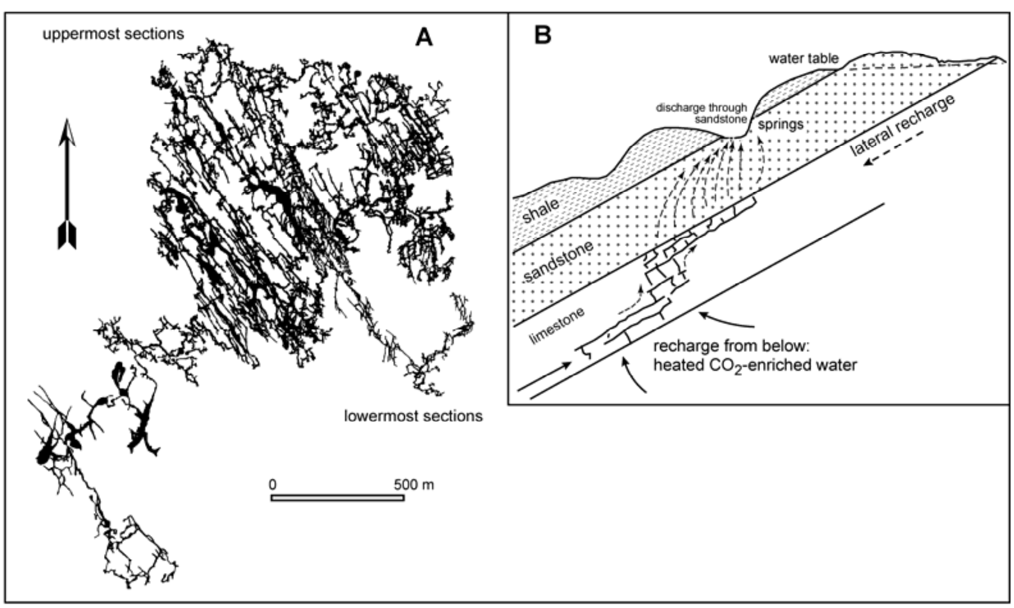

Figure 13. A = Plan of Wind Cave, Black Hills, South Dakota, USA (courtesy of Wind Cave National Park); B = a diagram illustrating the concept of the uplift origin of the Black Hills maze caves and the staircase effect of recharge-discharge offset. Redrawn from Ford (1989).

3.6 Dissolution processes in hypogenic speleogenesis

A common view, that confined conditions offer limited dissolution potential for karstification, is partially based on the deeply-rooted but generally inadequate simplistic concept of lateral flow through soluble formations, viewed as aquifers in an artesian basin. Alternatively, transverse hydraulic communication across lithological and stratiform porosity boundaries, which commonly coincide with major contrasts in water chemistry, gas composition and temperature, is especially potent in driving various disequilibrium and reaction-dissolution mechanisms.

Aggressiveness may represent an original undersaturation of groundwater with respect to the solid phase that is being entered, such as in the case of low-sulfate waters from underlying carbonates or sandstones entering a gypsum bed. Solubility of gypsum, quite high compared to carbonates, is further enhanced by several factors commonly operative in deep-seated environments:

- The solubility of gypsum increases by up to 3-6 times in the presence of various other dissolved salts, such as NaCl, $Mg(NO_3)_2$, etc. (the foreign ion effect);
- The reduction of dissolved sulfates removes sulfate ions from the solution and allows more sulfates to dissolve;
- Dedolomitization generates further dissolutional capacity with respect to gypsum, because Ca^{2+} is removed from solution and the sulfate ions react with Mg.

These processes also affect the origin of porosity in carbonates, so that interbedded evaporites and carbonates, a common stratigraphic arrangement in many basins, are particularly favorable for hypogenic speleogenesis (Palmer, 1995, 2000a).

Carbonic acid dissolution, which dominates overwhelmingly in epigenic carbonate speleogenesis, also operates as a hypogenic agent, though the origin of the acidity is different. It can be related to CO_2 generated from igneous processes, by thermometamorphism of carbonates, or by thermal degradation and oxidation of deep-seated organic compounds by mineral oxidants. The latter is common in hydrocarbon fields, where waters characteristically contain high CO_2 concentrations (Kaveev, 1963). Hydrogen sulfide is another common hypogenic source of acidity where there are sufficient sources of dissolved sulfate for reduction and where H_2S generated can escape from the reducing zones. However, it is generally believed that creation of significant caves, where these acids provide the dissolution mechanism, depends mainly upon rejuvenation of aggressiveness by mixing or cooling. These conditions are commonly met in the ascending limbs of intermediate or regional flow systems, when they interact with shallower flow systems.

Aggressiveness in hypogenic speleogenesis may also reflect acquisition of new sources of acid within the soluble formation itself, when groundwater flows transversely across it, or can be due to a number of mechanisms that rejuvenate the dissolutional capacity of fluids, such as the cooling of water, mixing of groundwaters of contrasting chemistry, and sulfate reduction and dedolomitization as mentioned above.

In shallower conditions, where H_2S-bearing waters rise to interact with oxygenated meteoric groundwaters, sulfuric acid dissolution can be a very strong speleogenetic agent. It is recognized as the main speleogenetic process for certain large caves (*e.g.* caves in the Guadalupe reef complex in the USA and Frasassi Cave in Italy) and many smaller caves. Based on this, some researchers distinguish sulfuric acid karst/speleogenesis as a peculiar type (Hill, 1996; 2000a; 2003a). Substantial sulfuric acid dissolution can also be caused by oxidation of iron sulfides such as pyrite and marcasite, where it is localized in ore bodies (Bottrell *et al.*, 2000). Lowe (1992) suggested that oxidation of pyrite along certain horizons or bedding planes in carbonates ("inception horizons") may create preferential flowpaths that later will be inherited by epigenic speleogenesis.

The increased solubility of calcite in cooling waters can cause dissolution along ascending flowpaths. The latter mechanism is commonly labeled as hydrothermal speleogenesis, occurring in high-gradient zones where ascending flow is localized along some highly permeable paths (Malinin, 1979; Dublyansky V., 1980; Dubljansky Ju., 1990, 2000a; Ford and Williams, 1989; Palmer, 1991; Andre and Rajaram, 2005). Solutional aggressiveness can be renewed or enhanced by mixing of waters of contrasting chemistry and dissolved gas content (Laptev, 1939; Bogli, 1964; Palmer, 1991), the effect widely cited in the karst literature as "mixing corrosion".

There are an increasing number of arguments and evidence suggesting that more than one process could be involved in many cases, operating either in combination or sequentially. Cross-formation flow is the main mechanism for hypogenic speleogenesis, which can integrate and trigger many cave-generating processes. Either carbonic acid or H_2S dissolution can operate in hydrothermal systems, which are essentially ascending transverse phenomena common in many basins. Mixing of contrasting waters is also commonly involved in hypogenic speleogenesis, at least at some stages. This, again, suggests that labeling types of karst and speleogenesis by a single dissolutional mechanism is misleading and should be avoided.

3.7 Mechanisms of hypogenic transverse speleogenesis

Field observations and numerous quantitative modeling studies (summarized in Palmer, Palmer and Sasowsky, 1999, and Dreybrodt, Gabrovšek and Romanov, 2005) suggest that speleogenesis in epigenic unconfined settings tends to produce broadly dendritic patterns of conduits due to highly competing development. Such development occurs because the positive feedback relationship between dissolution rate and discharge causes accelerated growth of selective favorable paths. Discharge increases with the growth of the conduit before and, more dramatically, after breakthrough. Discharge through a developing conduit is governed by the resistance of the conduit itself, by its narrowest downgradient part until the amount of available recharge begins to limit the flow. Another factor favoring the formation of branchwork patterns is that recharge in epigenic conditions is quickly rearranged in response to the competitive speleogenetic development, further concentrating flow through successful conduits.

In hypogene confined settings several important hydrogeologic and geochemical conditions account for the specificity of speleogenetic mechanisms involved, making them distinct from epigenic speleogenesis. The most important is that both recharge and discharge occur through adjacent insoluble formations or rocks of considerably lower solubility than a given cave formation, hence there is an external conservative hydraulic control on the amount of flow through the system. The resistance of the transverse conduit itself may control the amount of flow only in the very early stages of its growth. Then control is ultimately exerted by conductivity of the least permeable immediately adjacent formation, either the feeding or receiving one, or by the overall conductivity of the major confining formation. This suppresses the positive flow-dissolution feedback and hence speleogenetic competition in fracture networks, favoring the development of more pervasive conduit systems (maze patterns) where appropriate structural prerequisites exist. Palmer (2000b) has shown that discharge through the adjacent diffusely permeable formation (the feeding formation) is hardly affected by variations in width of the fracture being fed. Discharge to the fracture depends on the log of the fracture width, thus a ten-fold difference in fracture width produces only a two-fold difference in discharge. All fractures at the contact with the feeding formation receive nearly the same amount of recharge and grow at uniform rates.

Basic mechanisms of ascending transverse speleogenesis in a gypsum bed sandwiched between two non-soluble aquifers have been simulated by numerical modeling (Birk, 2002; Birk et al, 2003). The model combines a coupled continuum-pipe flow model, representing both diffuse-flow and conduit-flow components of karst aquifers, with a dissolution-transport model calculating dissolution rates and corresponding widening of karst conduits. Maze cave development is favored by the presence of systematic heterogeneities in vertical conductivity of a fracture network, which is shown above (Section 3.5) to be a familiar case because of discordance in the permeability structure between fracture networks at various intervals of a soluble formation. Hence, lateral components in speleogenetic development within certain beds are favored by the limited vertical connectivity of the adjacent fracture networks. In addition to structural preferences, the variation of boundary conditions in time, *e.g.* increasing hydraulic gradient across the soluble unit due to river incision into the upper confining bed, further influences the development of maze patterns.

Andre and Rajaram (2005) investigated dissolution of transverse conduits in hypogenic karst systems by rising thermal waters, using a coupled numerical model of fluid flow, heat transfer, and reactive transport. The key dissolution mechanism considered was the increased solubility of calcite along a cooling flow path. The physical domain of the model was a 500-m long fracture, with initial aperture of 0.05 mm and upward fluid flow at constant gradient. They found that during the very early stages of fracture growth, there is positive feedback between flow, heat transfer and dissolution. The period of relatively slow growth is followed by a short, abrupt period of rapid growth ("maturation" of Andre and Rajaram, an analogue to the "breakthrough" in the modeled development of early epigenic speleogenesis). However, soon after maturation, thermal coupling between the fluid and rock leads to negative feedback and a decrease in thermal gradient, especially near the entrance, resulting in shifting the growth area farther up into the fracture and in reduction of the overall fracture growth rate. They suggest that this suppresses the selectivity in conduit development in complex flow systems and allows alternative flow paths in a fracture network to develop, thus resulting in maze-like patterns.

These modeling attempts gave important insights into mechanisms of hypogenic transverse speleogenesis. However, they used highly simplifying assumptions for domain geometry and boundary conditions. More realistic fracture networks and more realistic positions of modeling domains within a regional flow system should be further studied, as well as effects of time-variant boundary conditions.

All numerical models studying the early evolution of dissolution conduits from initial fractures imply forced convection flow, and none of them account for buoyancy. It is commonly assumed that the buoyancy effects become significant when initial fractures are substantially enlarged. However, some recent studies demonstrate the pronounced role of buoyancy in rather small aperture fractures (Dijk and Berkowitz, 2000, 2002). In any case, it becomes more obvious now that buoyancy dissolution is of tremendous importance in hypogenic speleogenesis, which is strongly suggested by both regional karst and cave morphogenetic analyses (see next section and Section 4.3).

3.8 The role of free convection

Free convection (buoyancy) develops due to density variations in groundwater caused by solute concentration or temperature gradients, both being commonly pronounced in confined flow systems. Dissolution always tends to build up the solute concentration gradient that can drive buoyancy, hence free convection effects can be assumed to be inherently involved in speleogenesis, although they are overridden by forced convection in most unconfined systems. As forced-flow regimes in confined settings are commonly sluggish, the buoyancy-generating potential of dissolution itself gives rise to mixed convection regimes in most confined karst systems. In mixed-regime flow systems, speleogenesis commences in zones of upward cross-formational communication (Section 3.1), where the vertical hydraulic gradient and buoyancy potential are co-linear and congruous. The ascending transverse flow pattern is particularly favorable for density gradients of both types to develop (solute concentration and thermal) because less dense fluids enter a cave-forming zone from below, so that buoyancy effects play a particularly significant role in confined speleogenesis (Klimchouk, 1997b, 2000a).

Large-scale (regional) convection cells driven by density gradients are generally recognized to play a substantial role in groundwater circulation in sedimentary basins and adjacent massifs. The resultant flow patterns favor dissolution through various geochemical mechanisms. These aspects have been discussed in relation to many karst regions around the world. The effects are particularly pronounced in high-gradient zones of hydrothermal systems and where fresh groundwater comes in contact with evaporites from below.

Anderson and Kirkland (1980) provided a compelling demonstration that dissolution due to free convection is the main mechanism of ascending transverse speleogenesis in the thick sulfate and salt succession of the Castile and Salado formations in the Delaware Basin (western Texas and southeastern New Mexico, USA), resulting in the development of cavities at depth and cross-formational collapse breccia (breccia “pipes”) and masses of bioepigenetic calcite (“castiles” or “buttes”). Relatively fresh water is supplied through the underlying shelf and basin aquifers of the Delaware Mountain Group. Dissolutional chambers develop upward from the base of the Castile evaporites, with subsequent collapsing and formation of breccia. In other cases, the evolving transverse high permeability paths gave rise to replacement of sulfates by bioepigenetic calcite and the formation of “buttes” (Kirkland and Evans, 1980). The free convection flow pattern establishes itself in such zones, with rising fresh water and sinking dense brine components. The descending brine ultimately flows out through the basal aquifer to points of natural discharge.

A similar mechanism was shown to form vast chambers (Schlotten) in the upper Permian gypsum in the Sangerhausen and Mansfeld districts of Germany (Kempe, 1996). Schlotten are large voids, commonly isometric, elongated along the major tectonic fissures. About 100 cavities of this type are known in the region, encountered through the centuries in the course of mining operations at depths up to 400 m at the base of the Zechstein gypsum. Small cavities of this type, rising from the top of the underlying bed and remaining seemingly isolated from any integrated cave system, are commonly observed in gypsum quarries in the western Ukraine.

Speleogenesis at the base of deep-seated evaporites, driven by free convection from an underlying aquifer, is responsible for initiation of vertical breakdown structures called “breccia pipes,” “breccia chimneys,” “collapse columns,” “geologic organ pipes,” or “vertical through structures,” which are abundant in many deep-seated evaporite karsts as illustrated by many studies from the United States, Canada, China, Germany and Russia. They have diameters ranging from tens to over 100 m and propagate by upward dissolution and stoping from depths as great as 1200 m (Klimchouk and Andrejchuk, 1996).

Anderson and Kirkland (1980) generalized that *“If the source of the dissolving water is artesian or otherwise continuous, a flow cycle is developed in which the salt itself supplies the density gradient that becomes the vehicle of its own dissolution”* (1980, p.66). It can be further generalized that in confined settings the buoyancy “vehicle” operates in all major karst lithologies and at various scales, powered by solute concentration, thermal gradients, or both.

Curl (1966) provided a theoretical analysis of conduit enlargement by natural convection in a limestone aquifer, depicting transitional conditions that determine the prevalence of natural convection or forced flow regimes. He found that, with sufficiently slow water circulation (common in confined settings) convection caused by density differences might be the primary flow mode for limestone removal. This is made possible by even extremely small compositional differences. It is apparent that the effect becomes much stronger where evaporites are underlain by aquifers containing relatively fresh waters, and where thermal gradients are involved.

Particular morphological features resulting from free convection are discussed for limestone (hydrothermal caves), by Szunyogh, 1989; Bakalowicz *et al.* (1987), Collignon (1983) and Dublyansky (2000b, 2000c); for caves in gypsum by Kempe (1972, 1996), Klimchouk (1996b, 1997b, 2000a, 2000b); and in salt by Frumkin (1994, 2000). These features include cupolas, rising wall channels, ceiling channels and pendants, flat ceilings, inclined wall facets and keyhole cross-sections, although some of these features can be produced by other processes. However, when these forms occur in a characteristic suite, in which particular forms are in paragenetic relationships reflecting the operation of rising currents of dissolving water continuously from "feeders" to "outlets," they offer the most compelling morphological evidence for uprising flow patterns and the important role of free convection in hypogene transverse speleogenesis (see Section 4.3). Such suites are recognized in caves of many regions throughout the world, developed in evaporites and in carbonates.

Two types of free convection flow patterns can be recognized in hypogene speleogenesis (Klimchouk, 2000c). *Closed loop patterns* include rising limbs of less dense water and return (sinking) limbs of denser water, operating in the same pathway or segregated through adjacent alternative pathways. The solute load outflows via the same basal aquifer that supplies low-density water. *Open flow patterns* develop in mixed convection regime systems where there is an outflow pathway other than through the recharging aquifer, commonly an upper aquifer (above the soluble unit) and/or permeability structures that provide discharge to the surface through a leaky confining unit. Anderson and Kirkland (1980) described another variant of an open flow pattern in the Delaware Basin. Relatively fresh water is supplied laterally from the reef aquifer to the dissolving point of the dissolution "wedge" penetrating laterally deep through the evaporite succession. The dense brine produced by dissolution is drained downward into the lower aquifer, and ultimately flows out through it. For this pattern to operate, high permeability flow paths across the lower evaporite unit should have already been established by the upward-progressing dissolution (through the looped, or ascending open pattern), so the "descending" open cycle and the progression of a dissolution "wedge" probably come to operate during later stages of karst development. Given that buoyancy flow patterns are generally highly complex, both types of patterns may operate in a particular karst system.

It is commonly assumed that the buoyancy effects become significant when initial fractures are substantially enlarged. None of the numerical models studying the early evolution of dissolution conduits from initial fractures account for buoyancy. However, some recent studies demonstrate the pronounced role of buoyancy in rather small aperture fractures. Dijk and Berkowitz (2000, 2002) applied nuclear magnetic resonance imaging to quantitatively study the developing morphology, flow and dissolution patterns in natural, rough-walled, water-saturated halite fractures with 1-2 mm mean apertures. They found pronounced effects of buoyancy, resulting in vertical asymmetry during fracture evolution, with preferential dissolution at higher elevations. In horizontal fractures, the lower walls dissolve less rapidly than the upper walls (by a factor of ~2 to 3). The buoyancy effects in a vertical fracture with horizontal flow cause more rapid dissolution at higher elevation. For vertical fractures with upward flow, it is expected that the increasing saturation of the solution with elevation will be disturbed by the sinking, saturated solution. This will inhibit preferential dissolution at the lower (inflow) region and enhance dissolution farther downstream at higher elevations. Morphological studies in hypogene caves provide strong evidence for this effect.

4. Hypogenic cave features

Sedimentary basins and fold-and-thrust geological structures that experienced a variable degree of uplift and denudation, and contain carbonate and sulfate formations, are widespread throughout cratonic and fold-and-thrust regions. Most of them display features of ascending hypogene transverse speleogenesis, both relict (in the entrenched and drained sections) and currently operative (in confined, deep-seated sections, but also in entrenched settings). It is argued here that hypogene speleogenesis and resultant caves are much more widespread than commonly believed.

Identification and recognition of hypogenic caves is hindered and often confused due to three main causes. Firstly, as shown in Section 1.2, there is still considerable uncertainty in the scope of the hypogenic speleogenesis concept and in defining the term. Instead of restricting hypogenic speleogenesis to specific dissolutional mechanisms (hydrothermal or sulfuric acid), it is suggested here that hypogenic speleogenesis should be defined from the hydrogeological perspective, *i.e.* the formation of caves by recharge from below in leaky confined conditions by a number of dissolutional mechanisms that can be involved in specific cases. This approach is strongly corroborated by the remarkable similarity in morphological features exhibited by hypogenic caves formed by different dissolutional processes in different lithologies (Sections 4.2 and 4.4 below), by the overall regularities in the hydrostratigraphic occurrence of such caves, and by basin evolution analysis.

Secondly, the identification of hypogenic caves is difficult because, by the nature of their origin, they lack genetic relationships with the overlying or adjacent surface. They become accessible for direct observations when occasionally intercepted by denudational lowering, erosional entrenchment or human activities. However, the lack of genetic relationships with the surface may serve as one of the diagnostic criteria for hypogenic caves.

Thirdly, hypogenic caves become accessible being already relict, largely decoupled from their original cave-forming environment. Features created by epigenic processes often overprint original hypogenic morphologies, especially in wet climates, and sediment accumulations tend to mask important diagnostic features, especially in floors of caves.

Because of these difficulties, and also due to the still overwhelming dominance of epigenic concepts and models in karst science, many caves have been genetically misinterpreted. Even more known caves with “odd” patterns, which are not large or outstanding enough to attract specific scientific attention, remain without clear speleogenetic interpretation. Examples of regional speleogenetic analyses well connected with regional geological (hydrogeological and geomorphological) evolution are still scarce.

Speleogenetic considerations should be based on the broad evolutionary approach to karst development, as described in Chapter 2 and suggested by the classification of karst types presented in Figure 2. The inherent trend of basinal evolution during uplift is that deeper, confined, sections are being brought to the epigenic realm due to denudation, so that hypogenic caves pass through transitional conditions of initial breaching and draining, and get fossilized in the vadose zone. Hypogenic caves may pass the transient stages without major modification by the newly established unconfined flow patterns, or they may be considerably overprinted by epigenic processes. Where observed caves are in the transitional stages and overprint is obvious, it is most tempting to relate their origin to the contemporary epigenic conditions.

Hypogenic speleogenesis is largely independent of climate. Epigenic overprint over the inherited hypogenic cave features is particularly strong in regions of moderate and high runoff, from which most karst knowledge historically originated. In arid and semi-arid regions epigenic karst development and surface karst morphologies are normally subdued, but in the subsurface, hypogenic features are often abundant, resulting in a strong contrast in the degree of karstification between the surface and subsurface (Auler and Smart, 2003; Figure 14). The diagram in Figure 14 serves to clarify another common misconception about hypogenic karst in the karst literature, namely that hypogenic karst is peculiar to arid climates and less common in humid regions. In reality, it is just more readily recognizable in arid regions but it is overall equally present, although masked by epigenic development in shallow systems in humid karst regions.

Synthesis and generalization of available field data about hypogenic speleogenesis from various geological settings is crucial for more adequate interpretation of speleogenesis in particular cases, development of more adequate modeling scenarios, and for refinement and more adequate balancing of the theoretical basis of karst and cave science.

In identifying hypogenic caves, the primary criteria are morphological (patterns and meso-morphology) and hydrogeological (hydrostratigraphic position and recharge/flow pattern viewed from the perspective of the evolution of a regional groundwater flow system). The progress in understanding speleomorphogenesis in the hypogenic environment, and of regional hydrogeological regularities of hypogenic speleogenesis, makes the recognition of hypogenic caves easier and more reliable than before. Other criteria, such as mineralogical and sedimentological, are also important, although supplementary, and are useful for drawing inferences about particular geochemical dissolutional mechanisms and subsequent evolutionary history.

Various criteria, suggested earlier as diagnostic or indicative for either hydrothermal speleogenesis (Bakalowicz *et al.*, 1987; Dubljansky V., 1980; Dubljansky Ju., 1990, 2000b), sulfuric acid speleogenesis (Hill, 2000a; 2003a), or hypogenic speleogenesis viewed as specific only to hydrothermal and sulfuric acid dissolutional mechanisms (Palmer, 1991; Auler and Smart, 2003), are in fact pertinent to hypogenic transverse speleogenesis in the broader, hydrogeological sense, adopted here, regardless of the chemical mechanism involved.

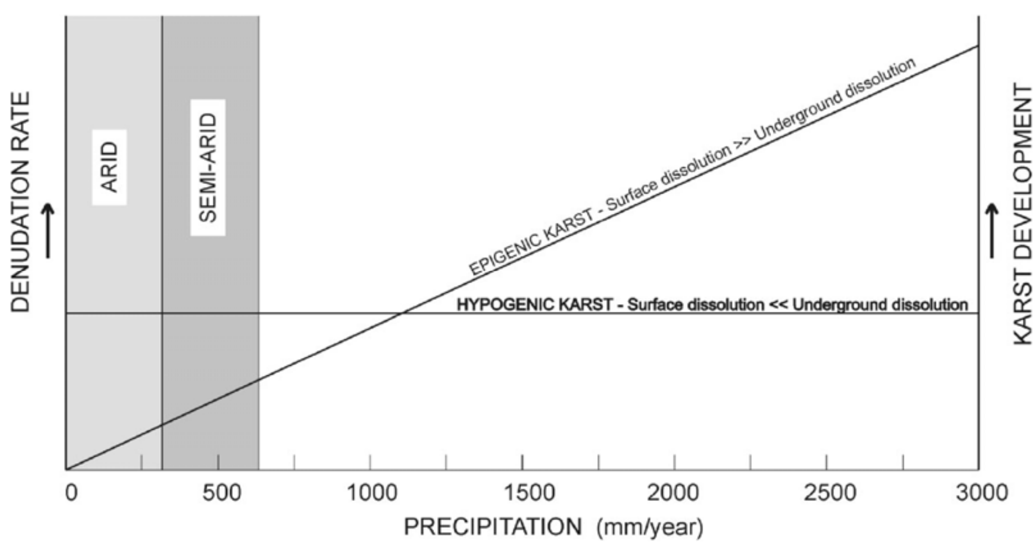

Figure 14. Schematic representation of the relative importance of hypogenic and epigenic components in shallow subsurface environments in different climatic settings (from Auler and Smart, 2003).

4.1 Criteria for distinguishing the hypogenic transverse origin for caves

The following geologic, morphologic, sedimentologic and mineralogic criteria are, in certain combinations, indicative of hypogenic transverse speleogenesis:

1. Presence of a source of recharge to the cave formation from below. It can be the immediately underlying aquifer, a laterally conductive bed within the aquifer system, or discrete vertical high-permeability paths conducting flow from still deeper aquifers. The common case is an insoluble porous or fractured bed, such as quartz sandstone or sand that serves as a regional aquifer and the source of water for transverse speleogenesis. It can also be a less soluble and more diffusely permeable material than the cave unit, such as oolitic limestone, densely fractured dolomite or marly limestone underlying gypsum or a less permeable limestone bed. To provide for dispersed and uniform recharge to the cave unit, the permeability structure of the source aquifer should be much more densely spaced than fissures in the soluble unit. Otherwise, discrete cavities would form in the cave unit, matching discrete paths of recharge from below.

2. Presence of an overlying aquifer bed. It can occur immediately above the soluble unit, or be separated by a thin leaky aquitard. The overlying aquifer acts as a governor for outflow, and allows transverse speleogenesis in a soluble bed to occur through areas offset from major flowpaths or breaches that discharge water out of the confined system. In some cases there can be no overlying aquifer, with a confining formation lying immediately above the cave formation. The confining formation should be considerably leaky to favor transverse speleogenesis in the cave formation.

3. Presence of a confining formation,commonly of regional extent and of low permeability. Transverse speleogenesis operates where the thickness of the confining strata is reduced due to erosional incision that induces considerable leakage, or where faulting or stratigraphic weaknesses allow discharge from the confined system to occur.

4. The overall layout of hypogenic cave systems and the position of their entrances show no genetic relationship to modern landscapes. However, significant cave development is normally induced by, and converges toward, valleys incising into the upper confining formation. Where modern valleys have incised below the cave-hosting formation, caves tend to border them. Paleovalleys, often buried, that cross modern watersheds could have induced transverse speleogenesis beneath them so that hypogenic cave systems can be found in the internal parts of modern intervalley massifs.

5. Cave patterns resulting from ascending transverse speleogenesis are strongly guided by the permeability structure in a cave formation. They are also influenced by the discordance of permeability structure in the adjacent formations and by the overall hydrostratigraphic arrangement (recharge-discharge configurations). Three-dimensional mazes with multiple stories or complex cave systems are common, although single isolated chambers, passages or crude clusters of a few intersecting passages may occur where fracturing is scarce (Section 4.2). Large rising shafts and collapse sinkholes, associated with deep hydrothermal systems, are also known.

6. Stories in three-dimensional mazes are guided by the distribution of initial porosity, which is commonly (but not always) stratiform. They may be horizontal or inclined, stratiform or discordant to bedding. Stories in ascending hypogenic systems form simultaneously within a complex transverse flow path, in contrast to epigenetic caves where stories reflect progressive lowering of the water table in response to the evolution of local river valleys, hence upper stories being older than lower.

7. When aggressive recharge from below is uniformly distributed, passages that hold similar positions in the system in relation to the flowpaths' arrangement (guided by the same set of fractures, or occurring within a single cave series or at the same story) are commonly uniform in size and morphology. A common feature of network mazes is high passage density. Spongework mazes may also occur where the initial porosity is represented by interconnected vugs. Larger volumes may be dissolved where aggressive recharge from below is concentrated by virtue of hydraulic properties of the feeding formation.

8. The characteristic features of ascending hypogenic cave systems are numerous blind terminations of passages in the lateral dimension and abrupt variations in passage cross-sections. Lateral changes indicate largely independent rising development of numerous almost independent transverse clusters (flow paths), and vertical changes indicate variations in initial porosity structures between lithological units.

9. The morphology of hypogenic caves, developed in varying lithologies by different dissolutional mechanisms, displays a very characteristic suite of similar forms indicating rising flow patterns during cave formation. This suite is the strongest diagnostic feature of hypogenic caves (Section 4.4) and consists of three major functional components: feeders (inlets), transitional wall and ceiling features, and outlet features.

10. Natural convection mechanisms (buoyancy-driven, upward pointed dissolution), powered either by thermal or solute differences, are widely operative in hypogenic caves, contributing significantly to the characteristic morphologies mentioned above and producing upward-directed flow markings (Section 4.4). Directional markings produced by vigorous flow regimes and lateral flow, *e.g.* scallops, are generally absent in hypogenic caves, although they may be present locally when considerable epigenic overprint occurs during the subsequent unconfined stage, *e.g.* by intercepted streams or backflooding. Water table markings, such as horizontal notches, may develop if the respective conditions are stable enough.

11. Clastic cave sediments are represented mainly by fine-grained clays and silts. These can be partly or largely autochthonous (comprising insoluble residues), although they often include considerable allochthonous sediments

brought into confined systems from overlying formations during the late artesian stages, mainly via breakdown structures.

12. Fluvial sediments are often absent but they can be present locally where invasion streams have superimposed onto a hypogenic system during exposure of the host formation and erosional entrenchment. Widespread deposition of backflooded silt sediments can occur during transitional stages.

13. Hypogenic caves are often barren of common infiltration speleothems unless the protective caprock is largely or entirely stripped. If the latter is the case, they may have abundant speleothems. "Exotic" minerals are often present, indicative of particular geochemical processes involved either in the cave formation or (particularly) in later transformations of geochemical environments during transition from confined to unconfined conditions. Hydrothermal minerals and minerals deposited as the products of redox reactions characteristic of transitional zones are common.

Although briefly summarized above, morphological features (cave patterns and cave mesomorphology) deserve more detailed consideration because they are most important for identifying cave-forming hydrogeologic environments, and thus the origin of their caves.

4.2 Cave patterns

Hypogenic caves display variable, often complex patterns. Branchwork caves, with passages converging as tributaries in the downstream direction, the most common pattern for epigenic speleogenesis, never form in hypogenic settings. Elementary patterns typical for hypogenic caves are *network mazes*, *spongework mazes*, *irregular chambers, isolated passages or crude clusters of passages,* and *rising shafts*. They often combine to form composite patterns, including complex 3-D structures. A variant of such complex patterns is distinguished as a *ramiform (ramifying) pattern,* which Palmer (1991, 2000a) described as "*caves composed by irregular rooms and galleries in a three-dimensional array with branches, that extend outward from the central portions*"". This description, however, is simply morphological and does not necessarily reflect organization (outward) of the cave-forming flow.

Network maze patterns are most common for hypogenic caves. Passages are strongly controlled by fractures and form more or less uniform networks, which may display either systematic or polygonal patterns, depending on the nature of the fracture networks. Systematic, often rectilinear, patterns are most common, reflecting tectonic influence on the formation of fracture networks. Polygonal patterns are guided by discontinuities of syndepositional or diagenetic origin. Examples include predominantly polygonal networks in the upper story of some western Ukrainian gypsum mazes, guided by early diagenetic structures (Figure 15) and a polygonal network of Yellow Jacket Cave in the Guadalupe Mountains, New Mexico, USA, guided by tepee-type syndepositional structures (Figure 16; see also Plate 16). Fracture and cave networks displaying different patterns may be present within a single area or at various stories/parts of a single cave, especially when confined to different rock units.

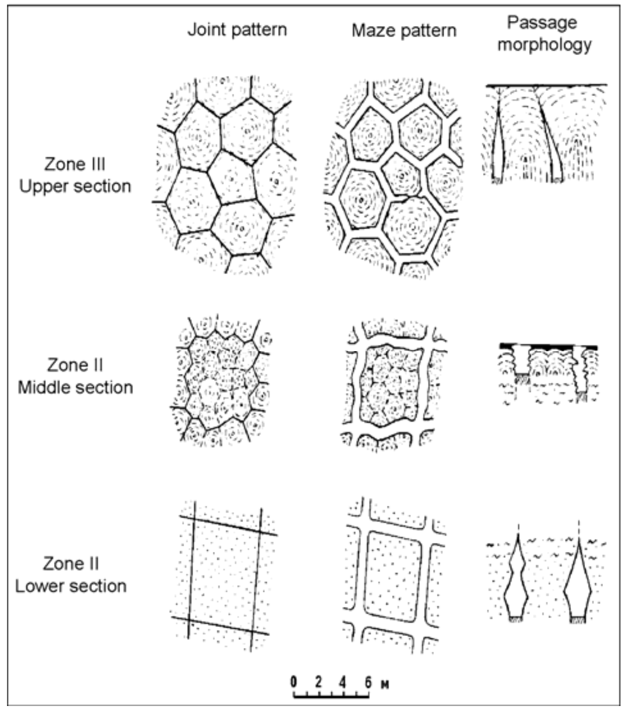

Figure 15. Variations in joint patterns and inherited maze patterns between different horizons of the Miocene gypsum bed in the western Ukraine, example from Optymistychna Cave. From Klimchouk et. al. (1995).

Spongework maze patterns are less typical than networks. Highly irregular passages develop through enlargement and coalescing of vuggy-type initial porosity in those horizons of the cave formation that have no major fractures but interconnected pores and vugs. Clusters or levels of spongework-type cavities are commonly combined with other patterns in adjacent horizons to form complex cave structures. An enlarged version of spongework, locally called boneyard, is represented in parts of some caves of the Guadalupe Mountains. In many hypogenic caves, it seems that rising buoyancy currents play a significant role in late stages of spongework development.

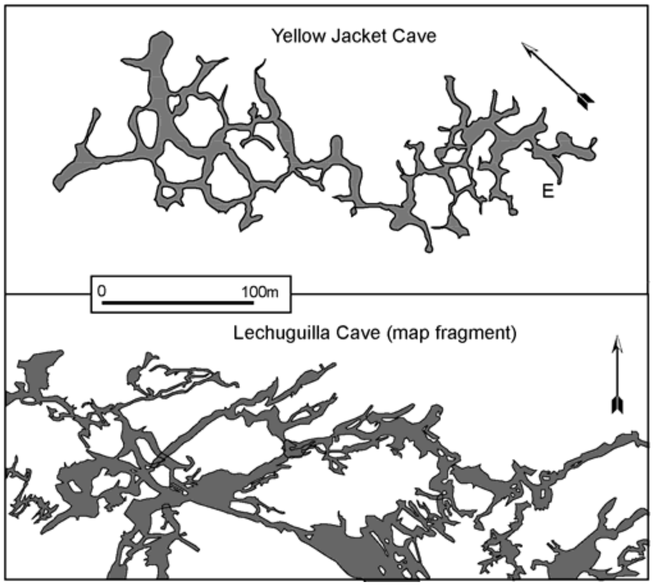

Figure 16. Examples of polygonal (above: Yellow Jacket Cave in the Yates Formation) and systematic (below: fragment of Lechuguilla Cave survey) network patterns in the same region, Guadalupe Mountains, USA. Yellow Jacket: simplified map from the original survey drawn by D.Belski, courtesy of the Pecos Valley Grotto. Lechuguilla: fragment from the original survey drawn by P.Bosted, courtesy of the US National Park Service.

Irregular chambers can be isolated cavities, or parts of composite patterns. In hypogenic settings they form in two situations: 1) by buoyant dissolution at the bottom of the cave formation, commonly evaporites, where a major aquifer immediately underlies it; 2) where the recharge from below is localized and flow and transverse speleogenesis in the cave formation is guided by prominent fractures. In the latter case, chamber development is commonly induced by intersection of the vertical flow path with a lateral flow-conducting horizon (stratiform permeability system) that enhances dissolution through mixing mechanisms. Irregular chambers in hypogenic karst can attain very large dimensions, such as directly documented cavities in evaporites of southern Harz, Germany (cavities of the "schlotten" type; Kempe, 1996), the Big Room in Carlsbad Cavern, or the indirectly documented (via drilling) hydrothermal cavity in the Archean and Proterozoic marbles in southern Bulgaria with a maximum vertical dimension of 1340 m and an estimated volume of 237.6 million m^3 (Sebev, 1970; Dubljansky,V. 2000), probably the largest known, although not accessible, cave chamber on Earth. It is likely that hypogenic megasinkholes associated with hydrothermal systems, such as Sistema El Zacatón in Mexico (Gary and Sharp, 2006; see Plate 19 and Section 4.5) or obruks (local name in Turkey for cenote-like sinkholes) in the Konya Basin, Turkey (Plate 19) are collapse features over giant chambers. However, it is also possible that they are rising dissolution shafts.

Isolated passages or crude clusters of passages also form in two situations: 1) in a manner similar to chambers, by buoyant dissolution at the bottom of the cave formation, where there is some initial linear guidance (by fractures or other kinds of weaknesses) but little or no forced flow across the formation; 2) by forced or mixed flow across a thin bed, where fracturing is scarce. In the former case some big irregular passage-like cavities may form, often associated with chambers, exemplified again by some "schlottens" in the South Harz (Kempe, 1996). In the latter case isolated slot-like passages or crude clusters of passages form, such as those intercepted by mines in the Neogene limestones in the southern Ukraine (see Figure 31).

Rising shafts are outlets of deep hypogenic systems and commonly hydrothermal. A type example is the 392 m deep Pozzo del Merro near Rome, Italy, presumably formed by rising thermal water charged with CO_2 and H_2S. It shows the morphology of a rising shaft (Figure 27), in contrast with the roughly cylindrical morphology of the sinkholes of Sistema El Zacatón in Mexico (see below), where hydrothermal cavities at depth are thought to open to the surface through collapse.

Consideration of hypogenic cave patterns only in plan view can be misleading, giving a false impression of seemingly two-dimensional structures in the case of laterally extensive network or spongework mazes. Further confusion arises from the fact that in many relict hypogenic mazes sediment fill obscures the "root" morphology at the passage floors, and that minor bottom features are rarely documented while surveying large maze caves even when they are recognizable. First recognized in the Western Ukrainian gypsum mazes and subsequently found in many maze caves around the world, both in limestone and gypsum, are numerous feeding conduits at lower levels, scattered throughout maze patterns at the master story (Figure 20; see Sections 4.2 and 4.4). With these feeders and their lower conduits, even largely horizontal laterally extensive maze patterns become complex three-dimensional structures.

Complex 3-D cave structures may develop within a rather thin formation (*e.g.* two- to four-story mazes in the western Ukraine confined within the 16-20 m thick gypsum formation) or extend through a vertical range of several hundred meters (*e.g.* Monte Cucco system, Central Italy: 930 m; Lechuguilla Cave in the Guadalupe Mountains, New Mexico, USA: 490 m). These complex 3-D structures often display a staircase arrangement of stories within a system, with cave areas at different stories shifted relative to each other (as discussed in section 3.5; see Figures 12 and 13), or have feeders at the lower level randomly or systematically distributed throughout the single master passage network. Some vertically extensive caves in the Guadalupe Mountains have prominent feeders as large isolated steeply ascending passages or clusters of rift-like passages connecting to some master level, and prominent outlet segments rising from the bulk of passages and rooms (Figure 17). These structures are composed of network and spongework mazes at various levels, coalescing with large chambers and passages and connected through rising vertical conduits. Other examples include the Monte Cucco system in Italy (Figure 24); complex bush-like upward-branching structures of hydrothermal caves in the Buda Hills, Hungary, composed of rising sequences of chambers and large spherical cupolas (Figure 22, B and C; Dubljansky Ju, 2000b); and network maze clusters at the base of the Joachim Dolomite in eastern Missouri, USA, with ascending staircase limbs of vertical pits and sub-horizontal passages (outlet component; Brod, 1964; Figure 39).

Multi-story mazes are variants of complex 3-D patterns. In a typical system, lower stories or individual rising conduits are recharge elements to a cave system. Master stories develop at intermediate elevations where there are laterally connected fracture systems. Upper stories serve as outflow structures ("outflow mazes" of Ford, 1989). Small patches of maze or lateral extensions of high cupola structures may develop at higher or highest elevations without bearing outflow functions ("adventitious" mazes of Ford), especially in systems where buoyancy flow plays a role.

In summary, the 3-D structure of hypogenic caves is controlled mainly by the distribution of initial permeability structures across the cave formation and adjacent formations, interaction of different permeability structures at various levels, and overall recharge/discharge conditions. Geochemical interaction of flow systems guided by transverse and lateral permeability pathways also may play a significant role. Buoyancy effects in free convection and mixed systems can also be important in creating complex cave structures.

4.3 The maze caves controversy

The most common (although not the only) pattern for hypogenic transverse speleogenesis is a network maze. Network mazes, often with several superimposed stories, constitute entire caves or parts of complex cave structures.

The formation of maze cave patterns has been specifically addressed in the karst literature for many years. Researchers who previously attributed the origin of maze caves to artesian conditions (*e.g.* Howard, 1964; White, 1969; Ford, 1971; Huntoon, 2000) or disregarded this possibility (Palmer, 1975, 1991, 2000b) implied the "classical" concept of lateral artesian flow through a soluble unit. Palmer examined the hydraulic-kinetic conditions within a simple loop in which water diverges into two branches that rejoin downstream, and showed that these branches cannot develop at comparable rates except at very high discharge to flow length (Q/L) ratios. Such conditions are not characteristic of lateral artesian flow, so he concluded that slow groundwater flow near chemical equilibrium, typical of confined aquifers, is least likely to produce maze caves (Palmer, 1975; 1991, 2000a).

White (1969) described the type of a "sandwich aquifer", where a thin carbonate unit is overlain and underlain by insoluble strata. He noted that network caves are characteristic for this situation and pointed out that such patterns form due to the lack of concentrated recharge from overlying beds.

Palmer (1975) specifically addressed the problem of maze patterns and suggested two main settings favorable for their development:

1) High-discharge or high-gradient flow during floods in the vicinity of constrictions in the main stream passages (floodwater mazes) and,

2) Diffuse recharge to a carbonate unit through a permeable but insoluble caprock such as quartz sandstone.

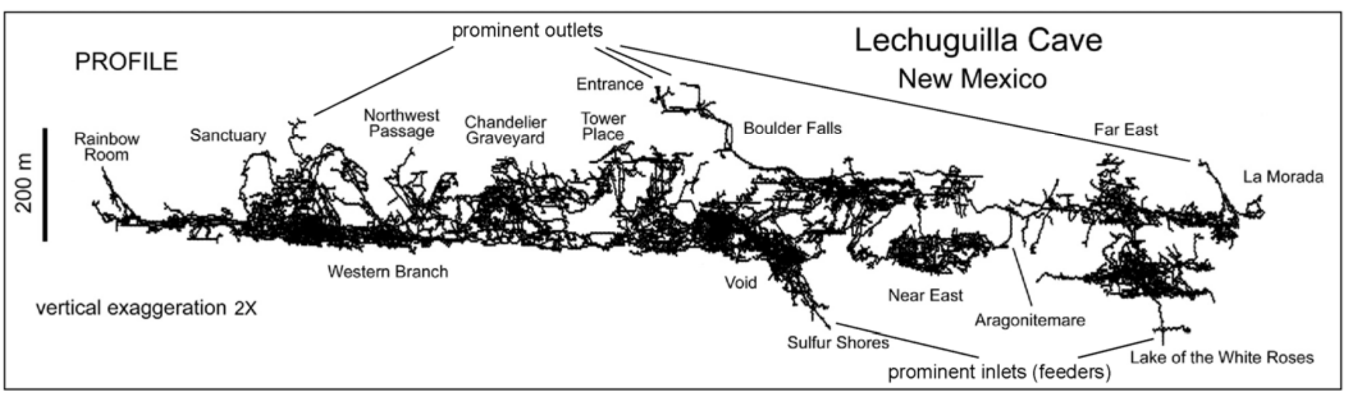

Figure 17. Profile of Lechuguilla Cave, NM, USA, by National Park Service volunteers, courtesy of US National Park Service. The cave is currently surveyed at 193.4 km in length and 490 m in depth. This is an example of a complex 3-D structure, in which some prominent inlet (feeding) and outlet components are easily recognizable.

Later he added the cases of sustained high gradients, such as beneath dams, and of mixing zones where the groundwater aggressiveness is locally boosted, and generalized that the formation of maze caves requires high Q/L ratios (Palmer, 2002).

Ford (1989) for the Black Hills caves and Klimchouk (Klimchouk and Rogozhnikov, 1982; Klimchouk, 1990, 1992; 1994) for the western Ukrainian caves suggested the model of maze development under confined conditions by dispersed ascending recharge from an underlying formation. Klimchouk (2000a, 2003a) generalized that this is the most common mechanism for confined speleogenesis.

An interesting suggestion of yet another mechanism of maze development is due to "phantomization" (rock-ghost weathering) by slow flow through fractures and dissolution of cement in the surrounding matrix at depth, followed by subsequent erosional removal of the impure residue in the vadose zone (Vergari and Quinif, 1997; Audra *et al.*, 2007a). As little is known about caves assigned to form by this mechanism, it is not discussed here.

A floodwater high gradient origin is a feasible mechanism for producing small mazes proximal to obstructions along well-defined stream passages conducting highly variable flow, or larger mazes in the epiphreatic zone of high-gradient alpine cave systems subject to quick and high rises of the water table (Audra *et al.*, 2007a). However, in relatively low-gradient environments (cratons and low mountains), it is less likely to create large maze clusters linked to rather small streams, such as in Skull Cave, New York, USA, often referred to as an example of floodwater development (Palmer, 2001; the cave plan on his Figure 10). An alternative possibility is that clusters of hypogenic transverse mazes, inherited from the confined stage, are encountered by invasion stream passages during the subsequent unconfined stage. A photograph of a typical floodwater (supposedly) passage on the cited figure shows a hole with a smooth edge in the bedrock floor, which is a typical example of a feeder (riser) in hypogenic transverse caves (see next section, photos F, H and I on Plate 3). Another frequently cited example of a floodwater maze is 21-km long Mystery Cave in Minnesota, USA, which is thought to form by the subterranean meander cutoff of a small river. The cave does function in this way at the present geomorphic stage, but recent examination of the cave by C. Alexander and the author revealed numerous morphologic features that strongly suggest a hypogenic transverse origin of the cave (see next section for discussion of hypogenic morphology and Section 4.5 for Mystery Cave). Meander cutoff flow has produced considerable morphological overprint and fluvial sedimentation in certain passages (Plate 14, upper left photo) but it has not erased hypogenic speleogenesis even in those central flow routes.

The floodwater model is often applied to explain mazes near rivers in somewhat static conditions (static "backflood mazes"). Although this might contribute to enlargement of already existing caves, it seems unlikely that mazes can *originate* in such situations because uniform early growth of initial porosity cannot be expected with side recharge and sluggish flow conditions, as shown with regard to lateral artesian flow (Palmer, 1975, 1991). Palmer's high Q/L ratio condition for maze development is not met in this situation. Furthermore, no maze caves referred to as being formed by backflood waters from the nearby river exhibit decrease in passage size or other regular changes in morphology in the direction away from the side recharge boundary, as would be expected if this origin were the case. Floodwaters from the nearby river can contribute to maze cave development where considerable conduit permeability is already available during river entrenchment, but it can be a self-standing speleogenetic mechanism only where there is intense open jointing.

The mechanism of diffuse recharge through a permeable but insoluble caprock, proposed by Palmer (1975) and widely used to explain maze patterns, requires additional discussion. It contains an important idea about the governing role of an adjacent porous formation for the amount of flow to fissures in a soluble unit (also expressed by White, 1969). This is the mechanism of restricted input/output that suppresses the positive flow-dissolution feedback and hence speleogenetic competition, as discussed in Section 3.7 in relation to confined settings and upward flow. However, the hydrogeological conceptual model that implies maze origin in unconfined settings by downward recharge from the overlying permeable caprock (Figure 18, A-B) has some problems if it is to be widely applicable. The hydrogeologic situation depicted represents certain evolutionary stages of breaching the caprock and the cave-hosting unit by denudation/erosion, and implies that it used to be a stratified multi-aquifer system, a common case in many sedimentary basins experiencing uplift and denudation. The model ignores the fact that flow in the low-permeability bed in this hydrostratigraphic setting (initially limestone) would be predominantly vertical, cross-formational, with descending flow within topographic/piezometric highs and ascending flow from underlying aquifers beneath valleys incising into the caprock. Most maze caves for which this origin was suggested are concentrated around river valleys or other prominent topographic lows (Palmer referred to diminished thickness of the caprock due to erosion; 2000b), which implies that ascending flow across the cave unit had been operative. Hence, such caves are fully compatible with the model of ascending (recharge from below) transverse speleogenesis.

From the perspective of basinal flow, as shown in Section 3.1, zones of ascending cross-formational flow in multi-aquifer systems are more important in supporting speleogenesis than zones of descending flow due to more vigorous circulation and a number of dissolution mechanisms that can be involved. It is significant that a hydrostratigraphic setting largely similar to that depicted by Palmer (1975; 2000b) was used to suggest the uprising development for maze caves (Ford and Williams, 1989; Figure 18-C), and this is, in fact, one of the basic settings discussed throughout this paper in the context of confined transverse speleogenesis.

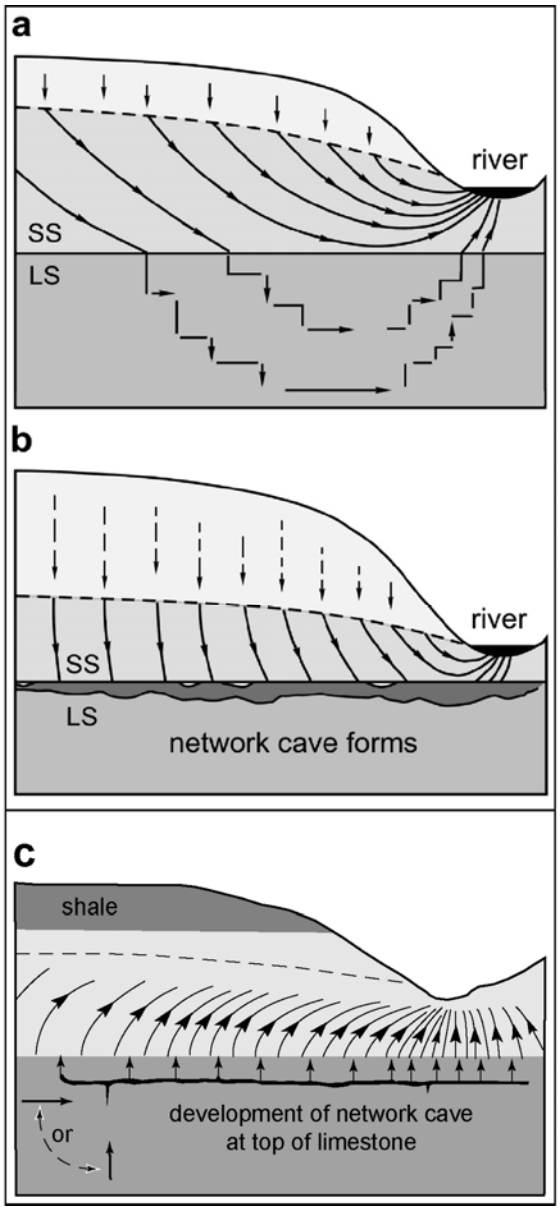

Figure 18. Conceptual models of development of a maze cave: A, B = by diffuse recharge from above (from Palmer, 2000b), C = by upward flow (from Ford and Williams, 1989).

Ford and Williams also pointed out that the Palmer model can explain only single-story mazes directly beneath the sandstone cover. It cannot explain development of multi-story mazes, and mazes occurring without direct contact with the bottom of the caprock, both being the most common cases of the structure and occurrence of maze caves.

Morphologically, there is no unambiguous evidence reported for maze caves that would suggest a descending flow pattern during their formation. Instead, at least in some of the caves referred to in various works to be formed according to this model, unambiguous evidence for the rising flow pattern has been recently recognized (the morphologic suite of rising flow; see next section). Eventually, those maze caves where a descending origin could be potentially supposed (where there is a permeable caprock currently exposed) are in all major respects similar to the caves where this origin can be definitely ruled out, *e.g.* beneath low-permeability cover, and where their confined transverse origin has been unequivocally established by bulk evidence.

A maze cave origin is frequently attributed to hydrothermal speleogenesis, the tendency reinforced by the paper by Bakalowicz *et al.* (1987), which suggested a hydrothermal origin for the Black Hills mazes. Other known examples of network mazes for which a hydrothermal dissolutional mechanism is well established are caves in the Buda Hills in Hungary. However, an emphasis on hydrothermal dissolution should not obscure the fact that these caves are attributed to a confined flow system and rising cross-formational flow, and that maze caves are known to form by a number of dissolutional mechanisms.

Frequent association of maze caves and hydrothermal systems can be easily explained by considering that deep basinal flow is commonly heated. Where structural and hydrodynamic conditions allow upward cross-formational flow, this commonly creates high-gradient thermal anomalies that favor hydrothermal dissolution. However, the origin of maze patterns is attributed not to hydrothermal dissolution (or to sulfuric acid dissolution, as some other work suggests) but to hydraulic conditions that favor disruption of the discharge-dissolution feedback mechanism. It was shown in Section 3.6 that a number of dissolutional mechanisms can operate in hypogenic transverse speleogenesis.

The broad evolutionary approach to speleogenesis implies that caves may inherit prior development through changing settings. Hence, the problem of cave origin requires specifying the mechanisms that were operative, and the features produced, during each of the main stages. The skeletal outline of a cave pattern is perhaps the most definite feature that can be attributed to certain recharge modes and flow systems (Palmer, 1991). As confined settings commonly pass into unconfined ones, phreatic through vadose, each subsequent setting may contribute substantially to cave development, sometimes adding significant volume to a cave. In this sense, both mechanisms questioned above with respect to the origin of maze caves, floodwater (backflooding) dissolution and dissolution by recharge from overlying permeable caprock, may certainly contribute to cave development, being operative during the respective transitional stages. However, the rapidly growing body of evidence from regions around the world (see Section 4.5) leads the author to believe that most known maze caves were formed in confined conditions, as the product of ascending hypogenic transverse speleogenesis.

4.4 Cave morphology

As with macro-morphological features (cave patterns), the meso-morphology of caves is the most important characteristic of caves indicating their origin. This is simply because a cave is primarily a form, produced by interaction between groundwater and its environment. The analysis of spatial and temporal relationships of different cave morphologies in the context of the regional geomorphic and hydrogeologic evolution is the most powerful tool for inferring cave origin.

As discussed in the preceding section, hypogenic caves may have variable patterns, controlled mainly by local geological and structural conditions (which also determine the mode of recharge from below). Despite this variability, meso-morphological features of hypogenic caves exhibit remarkable similarity among caves and comprise a specific set of forms.

Morphologic suite of rising flow (MSRF)

Some medium-scale morphological features of ascending hypogenic transverse caves have specific hydrologic functions and usually occur in a characteristic suite of forms; therefore they are particularly indicative of the mode of cave origin. Their occurrence in a suite makes the interpretation of their hydrologic function and origin especially unequivocal. Such a regular combination of forms, called here the *morphologic suite of rising flow* (MSRF), was first recognized in western Ukrainian gypsum mazes unequivocally established as ascending hypogenic caves, and subsequently found in many maze caves around the world in both limestone and gypsum. Some of the caves, where MSRF has been recognized, were previously attributed to hydrothermal, sulfuric acid or gypsum speleogenesis, or (maze caves) were viewed as developed by backflooding, by recharge through a permeable caprock, or not clearly interpreted genetically. The recognition of MSRF in such a great variety of caves, which have been previously seen as genetically different, suggests a common origin and is the strongest argument in favor of the dominant role of hydrogeological factors in speleogenesis, *i.e.* the type and regime of groundwater flow and the modes of recharge and discharge.

The morphologic suite of rising flow consists of three major components: 1) feeders (inlets), 2) transitional wall and ceiling features, and 3) outlet features (Figure 19).

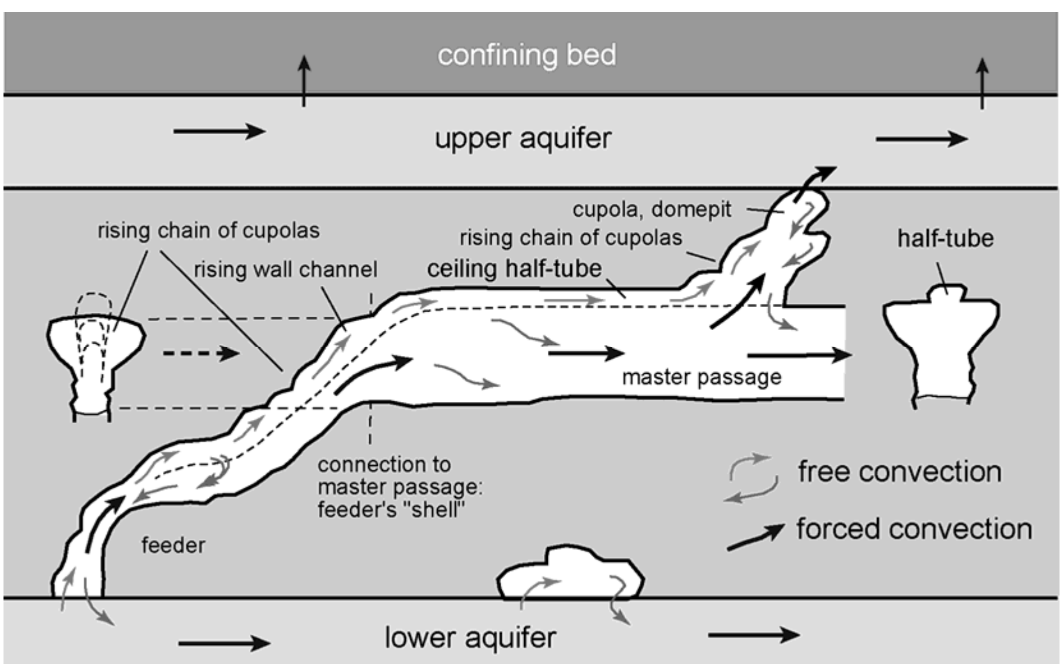

Figure 19. The morphologic suite of rising flow, diagnostic of confined transverse origin of caves. The geometry of a cave segment, the relative scale of features and hydrostratigraphy on this diagram is directly representative of Ozerna Cave in western Ukraine. However, the diagram is generic and elastic; it can be stretched vertically, and a complexity can be added to account for multiple lateral stories. The arrangement of the forms will repeat itself on each story, and functional relationships between the forms will remain the same.

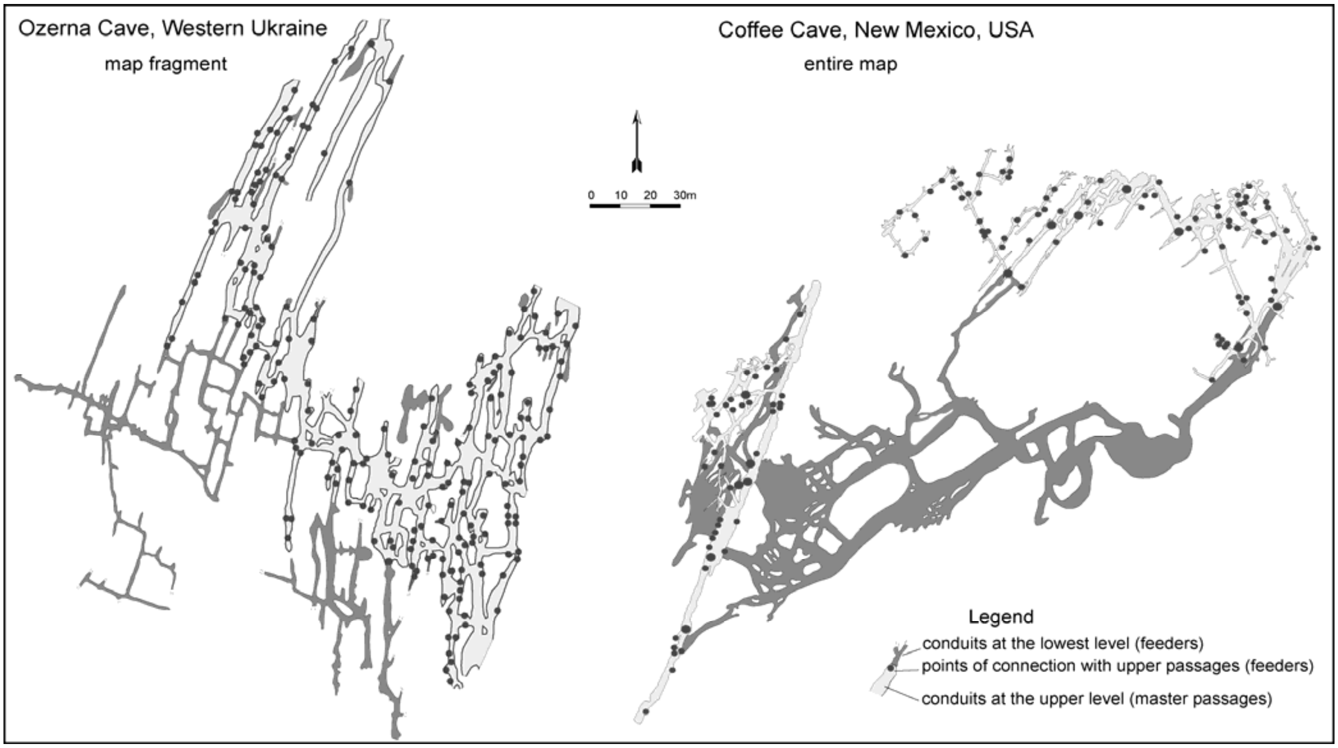

Figure 20. Distribution of point feeders (black dots; sub-vertical conduits connecting trunk passages) through the network of master passages in maze caves: Left = Ozerna Cave, western Ukraine (from Klimchouk, 1990); Right = Coffee Cave, New Mexico, USA (mapped and sketched by K. Stafford). Lower level passages locally form maze clusters that connect to the master level through sub-vertical point feeders.

1) *Feeders (inlets).* Original feeders are basal input points to hypogenic transverse systems, the lowermost components, vertical or sub-vertical conduits, through which fluids rise from the source aquifer. Such conduits are commonly separate but sometimes they form small networks at the lowermost story of a system, which bear the feeding function relative to the upper story. Feeders join master passages located at the next upper story and commonly scatter rather uniformly through their networks (Figure 20). Many feeders are point features; they may join the passage from a side (Plates 1 and 2), from the end (Plate 5, A, B and C), or scatter through the passage floor (Plate 3). Where master networks occur near the base of a soluble bed, they can receive recharge throughout the entire length of fissures to guide passage development. Feeders also can be rift-like features at the floor of master passages, which extend down to the contact with the underlying aquifer bed (Plate 5).

Master passages (in multi-story mazes) are passages that constitute laterally extensive networks within certain horizons of a soluble unit. They receive dispersed recharge from numerous feeding channels and represent the lateral component of flow due to discordance in initial porosity structure between different horizons. In some complex 3-D structures, there can be several stories of lateral development in the system. In that case, feeders of upper stories are the continuation of outlet features of the adjacent lower story. Hence, the lower stories function to recharge the upper stories. Sizes of feeders vary greatly, from small conduits (tubes, rift-like fissures, etc.) on the order of tens of centimeters to features many meters in diameter and tens of meters in vertical extent. In many instances, dimensions of feeder conduits are smaller in their lower parts and they often have ear-shaped orifices. This is due to buoyancy effects, shielding of walls in the lower parts from dissolution by more saturated water in the sinking limbs of free convection cells; and to mixing effects, which cause enhanced dissolution at the orifices due to mixing of waters of different chemistry.

Feeders are often obscured by the presence of sediment fill, but still can be identified in many cases by the presence of rising wall channels, or misinterpreted as “swallowing” or entrenchment forms rather than forms that transmitted rising flow.

2) *Transitional wall and ceiling features*. These features include rising wall channels, rising sets of coalesced ceiling cupolas or upward-convex arches, ceiling channels (half-tubes), and separate ceiling cupolas. They are usually arranged in continuous series, ultimately

connecting feeders to outlets, reflecting rising flow patterns and a considerable role of buoyancy effects (upward-focused dissolution by buoyant currents – rising limbs of free convection cells).

Rising wall channels (examples are on Plate 1) and rising sets of ceiling cupolas are found immediately adjacent to feeders, continued through ceiling half-tubes to cupolas and domepits. Rising sets of ceiling cupolas or series of upward-convex arches are also common for passages or rooms connecting different stories in a cave system (Plate 6, A through D).

Cupolas on the ceiling are commonly arranged in linear series comprising a kind of channel (Plate 6, H, I and K; Plate 7, B and C) but they can occur separately. In many cases where bottom features are observable, prominent cupolas or complex domes with numerous cupolas match in the plan view to particular feeders or groups of feeders at the floor, clearly suggesting a convection origin of the ceiling features. This origin for cupolas had been well recognized for hydrothermal caves (Müller and Sarvary, 1977; Dubljansky V., 1980; Lauritzen and Lundberg, 2000), but largely similar features at all scales are common for other types of hypogenic caves (sulfuric acid, "normal" limestone caves, caves in gypsum). Many cupolas have guiding fractures at their apexes but others show no such guidance. Cupolas alone are not exclusive to hypogenic speleogenesis; they may form in unconfined phreatic caves (reflecting the confinement of water within a passage itself), but their occurrence in a suite with other ceiling features as described here is clearly indicative of hypogenic speleogenesis and buoyant dissolution effects. Extensive discussion of cupolas has been recently provided by Osborne (2004).

Ceiling channels, also often called half-tubes, although commonly interpreted as paragenetic features formed when sediment fill directs phreatic dissolution upward, are very typical for hypogenic caves that have never been filled with sediment to the ceiling level. Instead, their relationships with feeders (through rising wall channels), and outlets in hypogenic caves, and rising patterns from the former to the latter, clearly suggest an origin due to buoyancy effects (Figure 19). In large passages or rooms where multiple feeders are present, several ceiling channels may braid in close proximity, leaving ceiling pendants in between. Particularly good examples of such pendants can be found in some gypsum caves in the western Ukraine, in the USA at Carlsbad Cavern, New Mexico, and in Caverns of Sonora, Texas. The vertical relief between pendants and adjacent channels can be as great as several meters, and such pendants are often well prepared to break down when a cave is drained and buoyant support is lost.

3) *Outlet features.* These are cupolas and domepits (vertical tubes) that rise from the ceiling of passages and rooms at a certain story and connect to the next upper story, or ultimately to the discharge boundary - the bottom of the overlying formation, a prominent bedding plane or the land surface. The ultimate outlets serve as discharge paths in a confined transverse system. Their ascending formation is suggested by their smoothed, curving walls, and by continuous morphology from connecting rising ceiling/wall features (Plate 8, A, B and D; Plate 9-I). In many caves, the bottom of the overlying aquifer bed ("receiving unit") is exposed at the outlet apex, sometimes with a gaping contact suggesting outflow via the bedding plane (Plate 10-A; see also Figure 9). Outlets that break into the next upper cave story, or to the ultimate discharge boundary are "successful" outlets, whereas blind-terminated cupolas can be regarded as "undeveloped" outlets. Closely spaced individual outlets in passages lying not far below the upper aquifer may merge to open the upper contact through a broader area along a passage (Plate 9-A), the ultimate case being where the upper contact is opened at the ceiling along the entire length of a passage (Plate 9-B).

Individual outlets can vary greatly in size, from less than a meter to many meters in cross-section and from less than a meter to tens of meters in vertical extent. Complex outlets from large systems may have composite morphology and rise for tens of meters from the main cave level (the entrance series of Lechuguilla Cave and the Spirit World above the Big Room in Carlsbad Cavern are good examples).

Plate 11 shows mega-outlets and gives an example of the described morphologic suite of rising flow, derived from the program presenting interactive 360^{o} panoramic views of Lechuguilla Cave, NM, created by Four Chambers Studio in collaboration with the US NPS. These views, with their three-dimensional range, proved to be a useful tool to study the cave morphology, enabling capture, even with certain skewing, of broad panoramas showing various morphologic components and their relationships.

Subaerial and other alternative possibilities for the origin of wall and ceiling features

Some individual morphs that compose the above-described suite were previously interpreted in different ways. See Ford and Williams (1989, 2007) and Lauritzen and Lundberg (2000) for overviews of cave meso-morphology, and Osborne (2004) for discussion of cupolas.

Cupolas (ceiling pockets) commonly occur in unconfined phreatic caves, reflecting the confinement of water within a passage itself. Such cupolas normally have simple forms and are not connected by ceiling half-tubes in

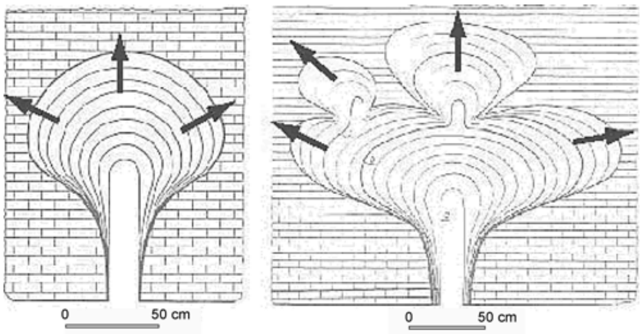

Figure 21. Theoretical development of simple and complex spheres by condensation corrosion. Simple sphere develops upwards and laterally (left); an irregularity in the ceiling produces a new sphere development that fits within the previous one (right); two neighbouring spheres diverge toward the greatest zone of heat transfer (right). After Szunyogh (1989) as adopted by Audra *et al.* (2007b).

this case; hence buoyant convection is not a contributing factor. Another interpretation is that cupolas form in subaerial conditions by moist air convection driven by the heat from a pool of thermal water in a chamber that is closed to outside airflow (Cigna and Forti, 1986). Acidic vapor (especially when H_2S is involved) is condensed onto cooler cave walls. Szunyogh (1989) showed that spherical pockets could be shaped in this way (Figure 21). Dreybrodt (2003), Dreybrodt *et al.* (2005) and Lismonde (2003) discussed complications with condensation processes arising from heat release at the wall surface, which slows down or terminates condensation. A sufficient gradient has to be maintained to enable continuing condensation. When this condition is met (*e.g.* above warm lakes in caves located close to the surface), the development of cupolas by condensation-corrosion (especially spherical and semi-spherical ones) is a sound possibility. It is likely that this process re-shapes original cupola-like forms created in confined/phreatic conditions. The condensation-corrosion mechanism does not serve to explain cupolas when they occur through all parts of extensive 3-D systems with vertical ranges of several hundred meters, including areas quite far from where warm lakes at the water table could be presumed (*e.g.* Monte Cucco in central Italy, Carlsbad Cavern and Lechuguilla Cave in the Guadalupe Mountains, USA, etc.).

There are additional arguments as to why the origin of cupolas by condensation-corrosion should not be applied too broadly. Cupolas are common in hypogenic caves for which neither thermal nor sulfuric acid processes are applicable, such as hypogenic caves in gypsum. Among caves whose origin involved hydrothermal or/and sulfuric processes, cupolas are common also in those where no signs of water table effects are recognizable, such as maze caves composed of small passages arranged in inclined stories. In 3-D cave systems, cupola/domepit complexes often extend upward from a base passage or chamber for tens of meters and are terminated at, or interrupted by, differently oriented stories of maze passages with which the cupola/domepit complexes show clear functional relationships. Their development due to condensation-corrosion seems to be highly unlikely in such situations. In a broader context, the water table model of hypogenic speleogenesis is discussed in Section 4.5 in relation with the Guadalupe Mountains speleogenesis.

Traditional interpretation of ceiling half-tubes (ceiling channels) is that they are paragenetic features formed when sediment choke of passages directs phreatic dissolution upward (Ford and Williams, 1989; Lauritzen and Lundberg, 2000). This is an obvious case in many epigenic caves. However, half-tubes are commonly observed in caves that have never been filled by sediments, such as in most hypogenic caves. Their incompatibility with the paragenetic model is especially evident in multi-story and complex 3-D caves where half-tubes occur at different levels, are connected by rising forms from below, and connect to cupolas/domepits at higher ceiling elevations.

Pendants are residual pillars of rock between channels cut into the ceiling. They are traditionally interpreted as remnants of bedding plane anastomoses (when the main body of passages had entrenched down) or as pillars between closely-spaced paragenetic ceiling channels. This fits well to observations in many epigenetic caves. However, both explanations are not applicable to many hypogenic caves where broadly braiding ceiling channels (creating pendants in between them) best fit to the model of buoyancy currents rising from multiple feeders at the bottom.

Rising wall and ceiling channels are sometimes explained as trails curved by degassing bubbles in phreatic thermal CO_2-H_2S systems (Audra *et al.* 2002). However, Palmer and Palmer (2000a) noted that the maximum depth at which degassing to form bubbles can take place is limited to a few meters below the water table at commonly observed concentrations of these gases. Rising channels are widespread in hypogenic caves at all levels within vertically extended caves. Furthermore, similar rising channels are common in gypsum caves and caves where degassing of rising water and enhanced condensation-corrosion cannot take place.

Widespread occurrence of the above features in the characteristic suite of forms (which also includes feeders), in a variety of caves that share a hypogenic origin in the hydrogeological sense, strongly suggests their interrelated origin as described in the previous section and depicted in Figure 19. In hypogenic transverse systems, local convection cells can develop from even small density gradients (either thermal or solute) in mature caves under confined conditions of sluggish rising forced flow and homogenous hydraulic heads. Less dense and more aggressive water tends to occupy the uppermost position in the available space geometry, producing upward-directed imprints such as rising wall channels, ceiling half-tubes, and cupolas. Buoyancy currents begin from feeders – points from which water entered a cave or a particular story. Buoyant dissolution morphologies comprise a continuous series, well recognizable in caves where the original morphology was not much disrupted or obscured by later water table and vadose development, breakdown processes, or sedimentation. The morphologic suite of rising flow is best represented in limestone caves where thermal waters were involved, and in gypsum caves where the gypsum strata are underlain by an aquifer with relatively low solute load.

Dead ends, abrupt changes in morphology and partitions

Some morphologic features in caves, such as blind terminations of passages (dead ends), abrupt changes in size and morphology, and various kinds of bedrock partitions (vertical or horizontal) were always regarded as odd and puzzling by researchers accustomed to "lateral" speleogenetic thinking. They are difficult to explain within the conventional speleogenetic concepts of caves formed by lateral flow or by dissolution at the water table. These features are sometimes considered as attributive to sulfuric acid speleogenesis (*e.g.* Hill, 2003a, 2006, Hose and Macalady, 2006) but in fact, these are very common for most hypogenic caves regardless of the dissolution chemistry involved and host rock composition. These features are perfectly consistent with rising transverse speleogenesis; lateral changes simply indicate largely independent rising development of numerous transverse segments (flow paths), and vertical changes indicate variations in initial porosity structures across a vertical section.

Blind terminations of passages are inherent elements in almost all maze caves (see cave maps throughout this book) and complex 3-D caves. In most cases they are "dead ends" only from a "lateral" perspective but in the transverse flow scheme they are open either to recharge (feeders from below; Plate 2, A-D; Plate 5, A-C) or to discharge (outlets to above). The transverse speleogenesis mechanism allows even a single, laterally isolated fracture to enlarge to a passable size by vertical flow through its entire length, but the passage will remain blind-terminated (pinching out) laterally at both ends (Figure 31-A).

Partitions are thin separations between adjacent passages or chambers made up of bedrock or various kinds of planar resistant structures exhumed by dissolution, such as lithified fill of fractures or faults and paleokarstic bodies. They are common in many densely packed maze caves, where bedrock separations between passages are commonly thin (Plates 12 and 13). In the Western Ukrainian mazes, bedrock separations ("pillars") between adjacent passages may be less than a meter thick (Plate 12, C though G). Sometimes they are only a few centimeters thick so that a "window" can be broken by a punch. When water table overprint was locally noticeable on transitional stages, thin partitions can be easily truncated by dissolution at the water table (Plate 12, E-G).

Another type of partition is represented by projections of lithified fracture fill exposed by dissolution. They may largely or completely partition rather large passages (Plate 13). Common in some mazes of the western Ukraine, such partitions are quite fragile (being only a few centimeters thick). The fact that they remain intact, and passage morphology remains uniform on both sides of such partitions, indicates a homogenous head field within a mature cave system and an overall transverse flow pattern. Horizontal partitions by more resistant beds in a stratified sequence may create multi-story cave systems, where passages of different stories are closely spaced in a vertical cross-section (*e.g.* Endless and Dry caves in the Guadalupe Mountains, New Mexico, USA; Archeri Cave in the Minor Caucasus, Armenia; Coffee Cave in the Roswell Basin, New Mexico, USA, Stafford *et al.*, 2008). Osborne (2003) described partitions of various kinds in Australian caves and recognized that caves containing vertical and sub-vertical partitions are likely to be formed by *per ascensum* speleogenetic mechanisms.

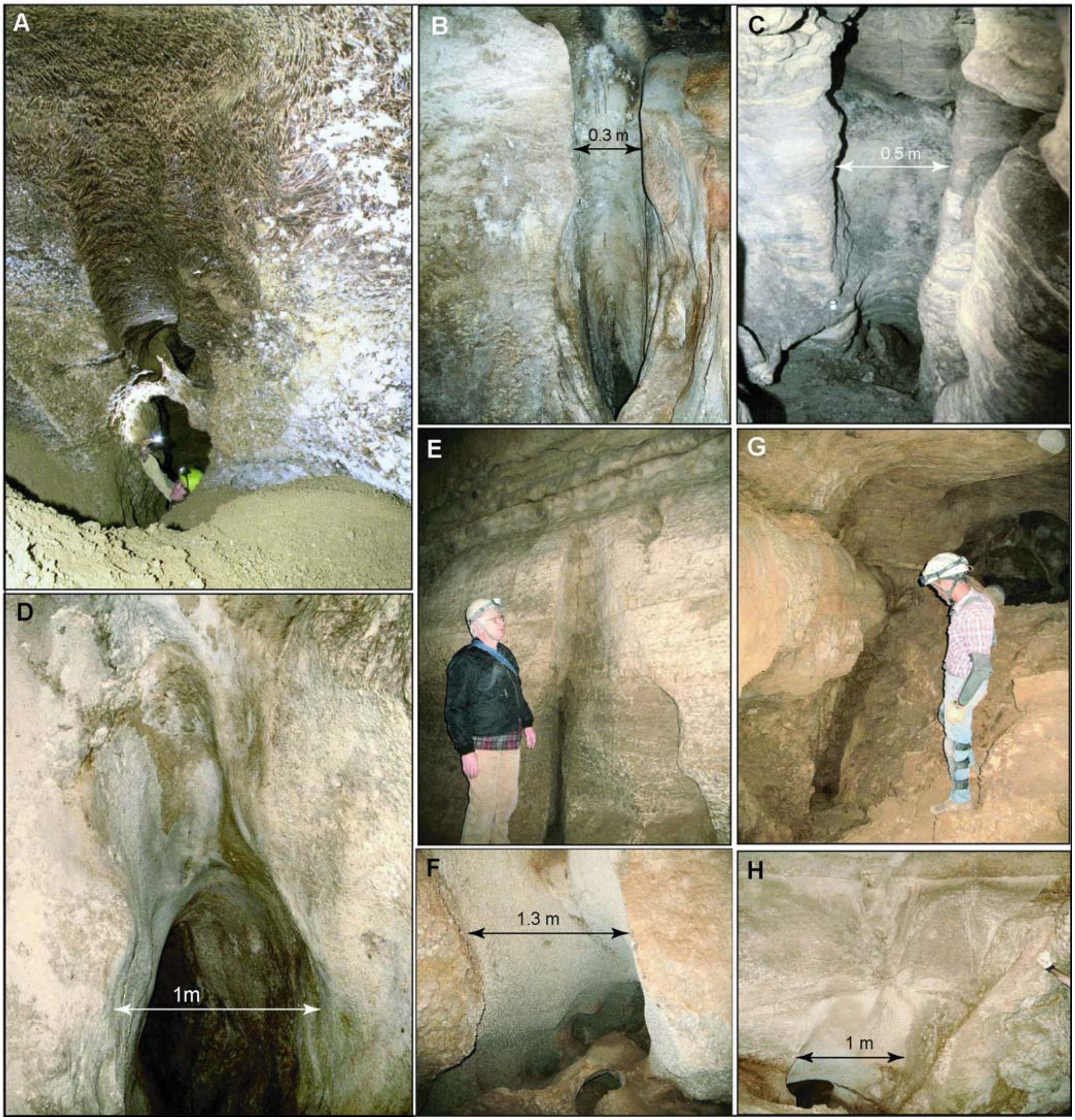

Plate 1. Feeders: side feeders with rising wall channels. A = Optymistychna Cave, western Ukraine (Miocene Gypsum); B, D = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); C = Fuchslabyrinth Cave, Germany (Triassic Muschelkalk limestone); E = Mystery Cave, MN, USA (Ordovician limestone); G = Dry Cave, Guadalupe Mountains, NM, USA, (Permian limestone); F, H = Deep Cave, TX, USA (Cretaceous limestone). Photos by A. Klimchouk.

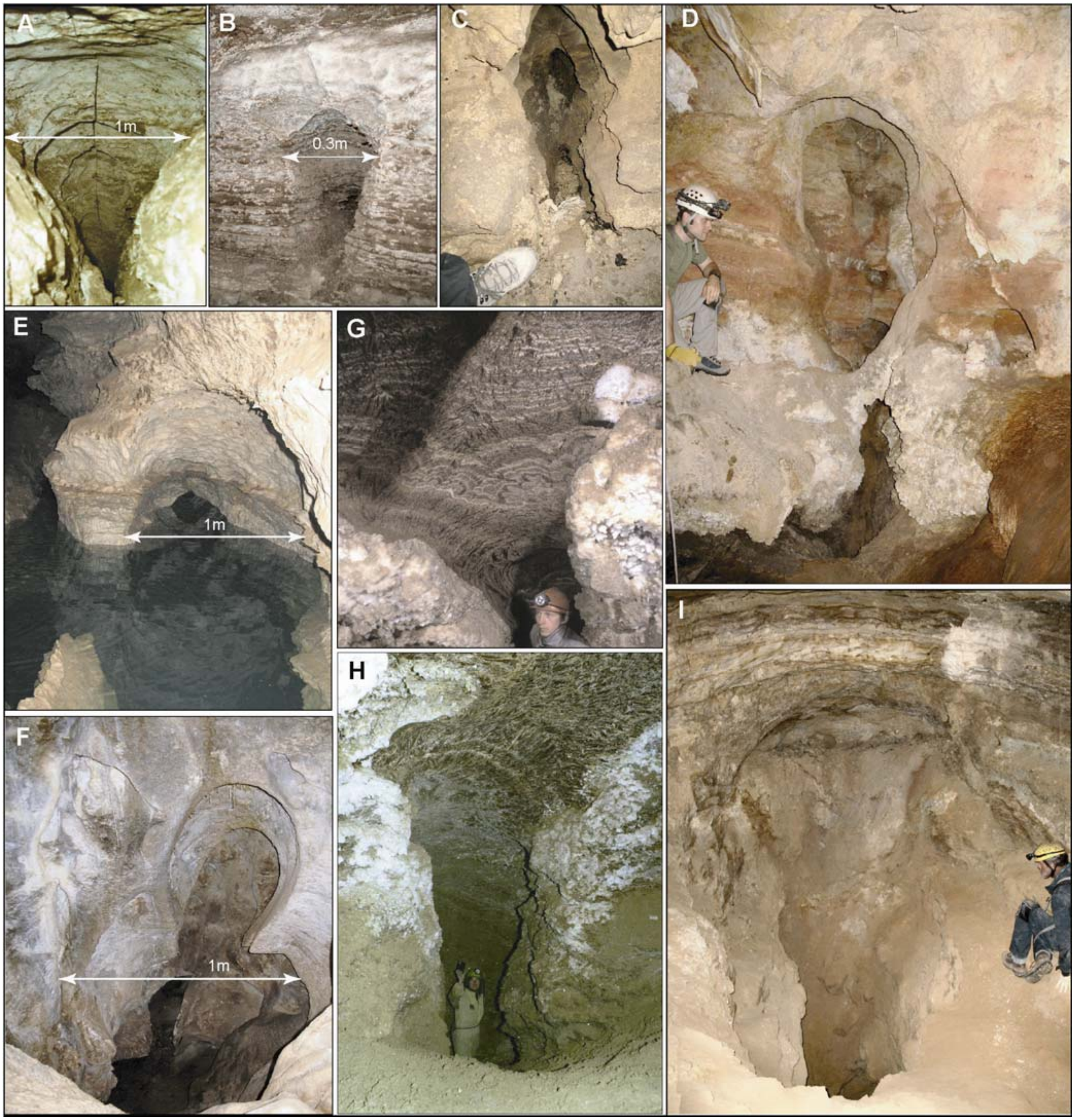

Plate 2. Feeders: side feeders with ear-like or domed orifices. A = Ozerna Cave, western Ukraine (Miocene gypsum); B, E = Coffee Cave, NM, USA (Permian gypsum); C = Robber Baron Cave, TX, USA (Cretaceous limestone); D = Spider Cave, NM, USA (Permian limestone); G = Dzhurinskaja Cave, western Ukraine (Miocene gypsum); F = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); H = Optymistychna Cave, western Ukraine (Miocene gypsum); I = Endless Cave, NM, USA. Photos by A. Klimchouk.

Note: A feeder in Endless Cave, Guadalupe Mountains, NM (photo I), is rimmed with a massive deposit of endellite (hydrated halloysite), a clay alteration mineral indicative of sulfuric acid. Other occurrences of endellite rimming feeders are found in Amazing Maze Cave in Texas (see Plate 4-D).

Plate 3. Feeders: point features in passage floors. A = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); B, C, D and I = Blowing Hole, Florida, USA (Eocene limestone); E = Mlynki Cave, western Ukraine (Miocene gypsum); G = Dry Cave, Guadalupe Mountains, NM, USA (Permian limestone); F = Robber Baron Cave, TX, USA (Cretaceous limestone); H = Fuchslabyrinth Cave, Germany (Triassic Muschelkalk limestone). Photos by A. Klimchouk.

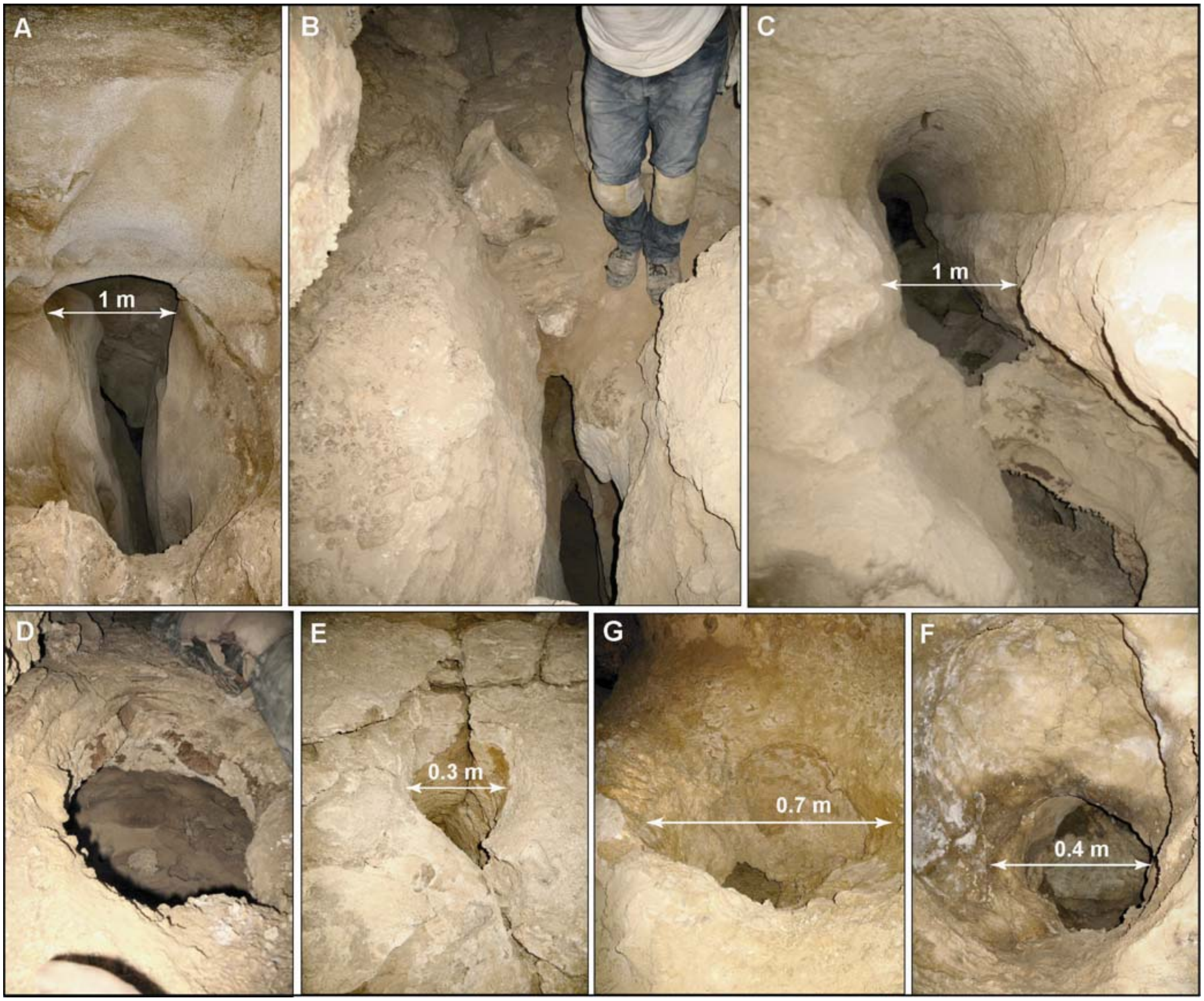

Plate 4. Feeders: point features in passage floors and lower walls. A = Deep Cave, TX, USA (Cretaceous limestone); B, C, D and E = Amazing Maze Cave, TX, USA (Cretaceous limestone); G = Caverns of Sonora, TX, USA (Cretaceous limestone); F = Dry Cave, NM, USA (Permian limestone). Photos by A. Klimchouk.

Note: A feeder in Amazing Maze Cave, TX (photo D), is rimmed with a massive deposit of endellite (hydrated halloysite), a clay alteration mineral indicative of sulfuric acid. Other occurrences of endellite rimming feeders are found in Endless Cave, Guadalupe Mountains, NM (see Plate 2-I).

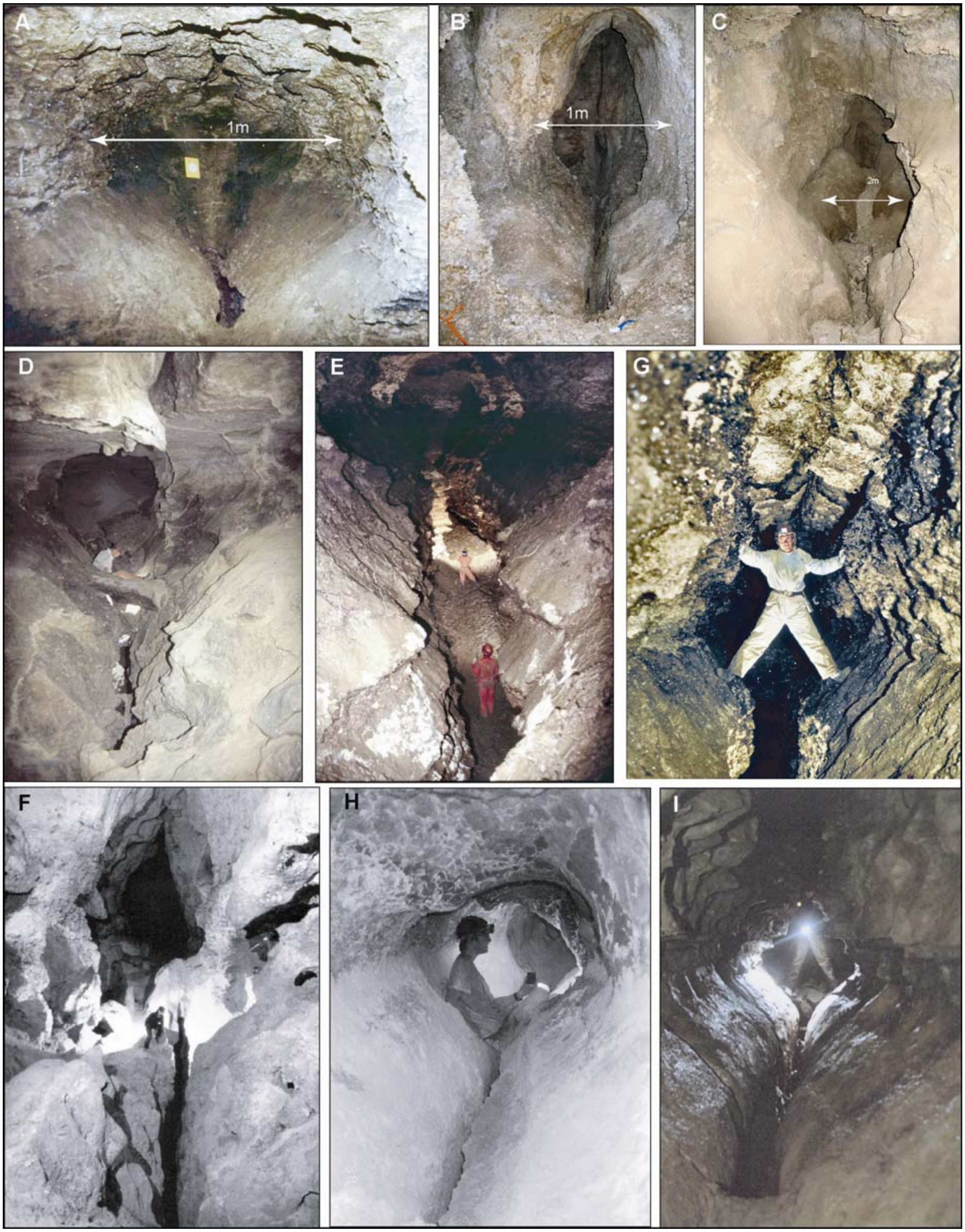

Plate 5. Feeders: fissure- and rift-like feeders at in passage floors. A, B and C = feeders in dead ends, A = Mlynki Cave, western Ukraine (Miocene gypsum); B = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); C = Coffee Cave, NM, USA (Permian gypsum); D = Mlynki Cave, western Ukraine (Miocene gypsum); E = Zoloushka Cave, western Ukraine (Miocene gypsum); G = Ozerna Cave, western Ukraine (Miocene gypsum); F = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); H = Aneva Cave, Israel (Cretaceous limestone); I = Knock Fell Caverns, Northern Pennines, UK (Carboniferous limestones). Photo E by V. Kisseljov, photo F by A. Palmer, photo H by A. Frumkin, other photos by A .Klimchouk.

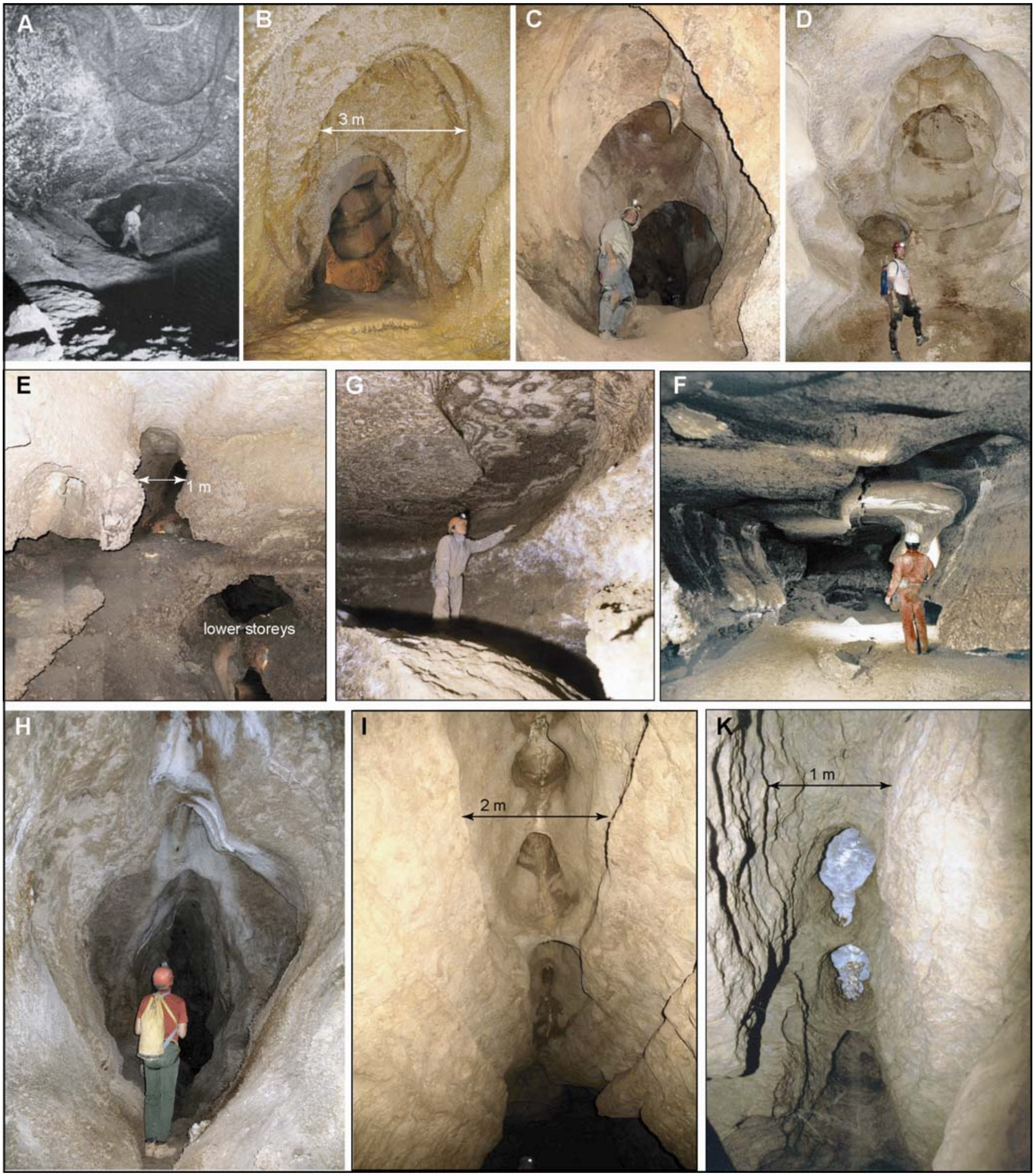

Plate 6. Rising chains of ceiling cupolas and upward-convex arches (A through D), ceiling channels ("half-tubes"; E through F) and serial cupolas in the ceiling apex H through K). A = Optymistychna Cave, western Ukraine (Miocene gypsum); B and E = Caverns of Sonora, TX, USA (Cretaceous limestone); C = Spider Cave, Guadalupe Mountains, NM, USA (Permian limestone); D = Deep Cave, TX, USA (Cretaceous limestone); G and F = Dzhurinskaja Cave, western Ukraine (Miocene gypsum); H = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); I = Mystery Cave, MN, USA (Ordovician limestone); K = Fuchslabyrinth Cave, Germany (Triassic Muschelkalk limestone). Photos by A.Klimchouk.

Plate 7. Ceiling channels and cupolas in linear series. A = Atlantida Cave, western Ukraine (Miocene gypsum); B = Carlsbad Cavern, Guadalupe Mountains, NM, USA (Permian limestone); C = Mystery Cave, MN, USA (Ordovician limestone); D = Lechuguilla Cave, NM, USA (Permian limestone). Photo D by S.Allison, other photos by A.Klimchouk.

Plate 8. Outlets with connecting ceiling channels (A, B and D) or formed within a ceiling channel, viewed from below. Arrows on solid lines give a scale (approx. 1 m). Dashed lines indicate direction of the rising ceiling and reflect rising buoyancy currents. A, B, C and D = Carlsbad Cavern (Permian limestone), Guadalupe Mountains, NM, USA; E = Dry Cave, Guadalupe Mountains, NM, USA (Permian limestone); F = Blowing Hole, Florida, USA (Eocene limestone). Photos by A.Klimchouk.

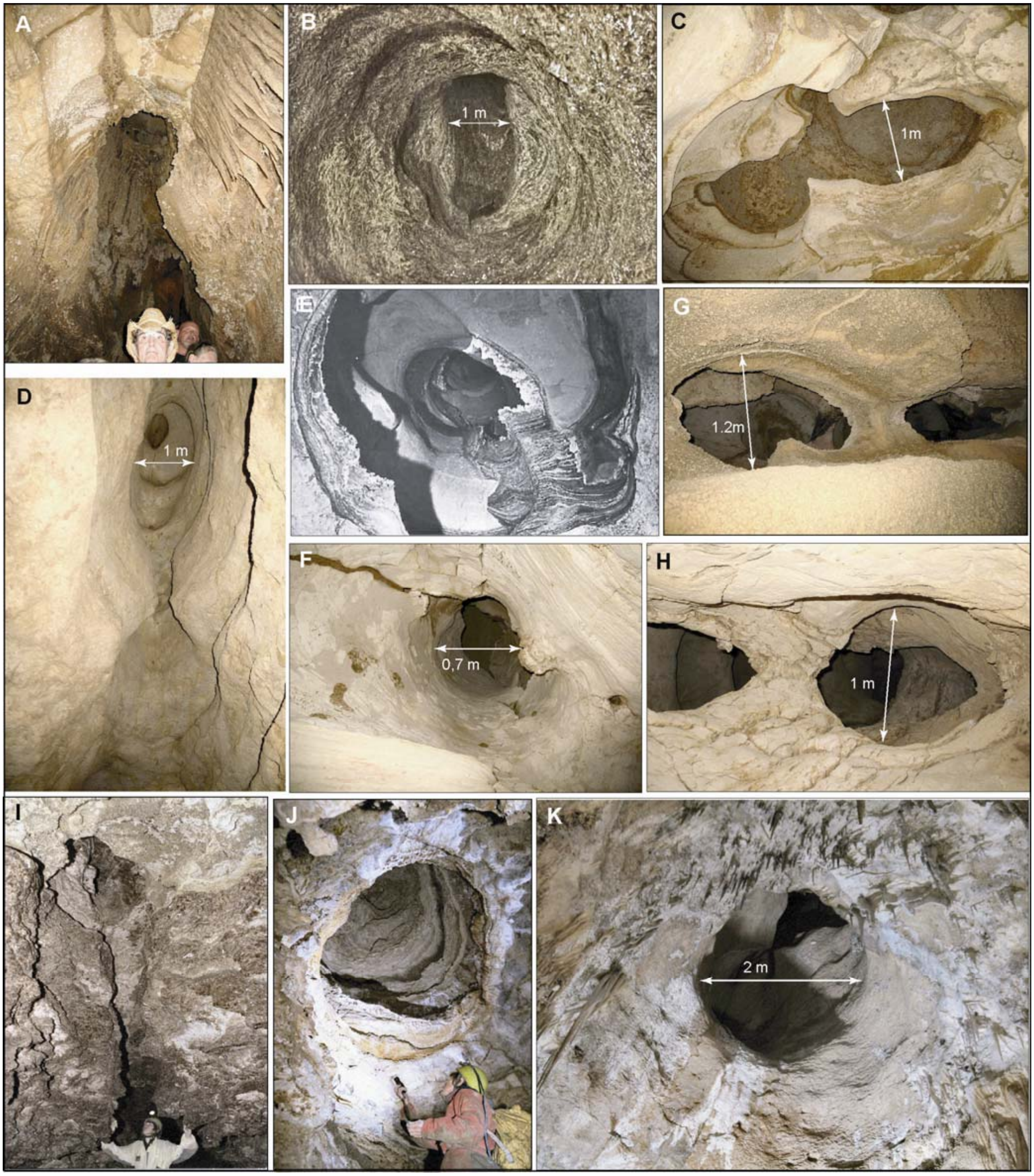

Plate 9. Outlets in cupolas and domepits, viewed from below. A, C and G = Caverns of Sonora, TX, USA (Cretaceous limestone); B = Optymistychna Cave, western Ukraine (Miocene gypsum); D = Mystery Cave, MN, USA (Ordovician limestone); E = Wind Cave, SD, USA (Carboniferous limestone); F and H = Amazing Maze Cave, TX, USA (Cretaceous limestone); J = Slavka Cave, western Ukraine (Miocene gypsum); K = Carlsbad Cavern (Big Room), NM, USA. Photo E by A.Palmer, other photos by A.Klimchouk.

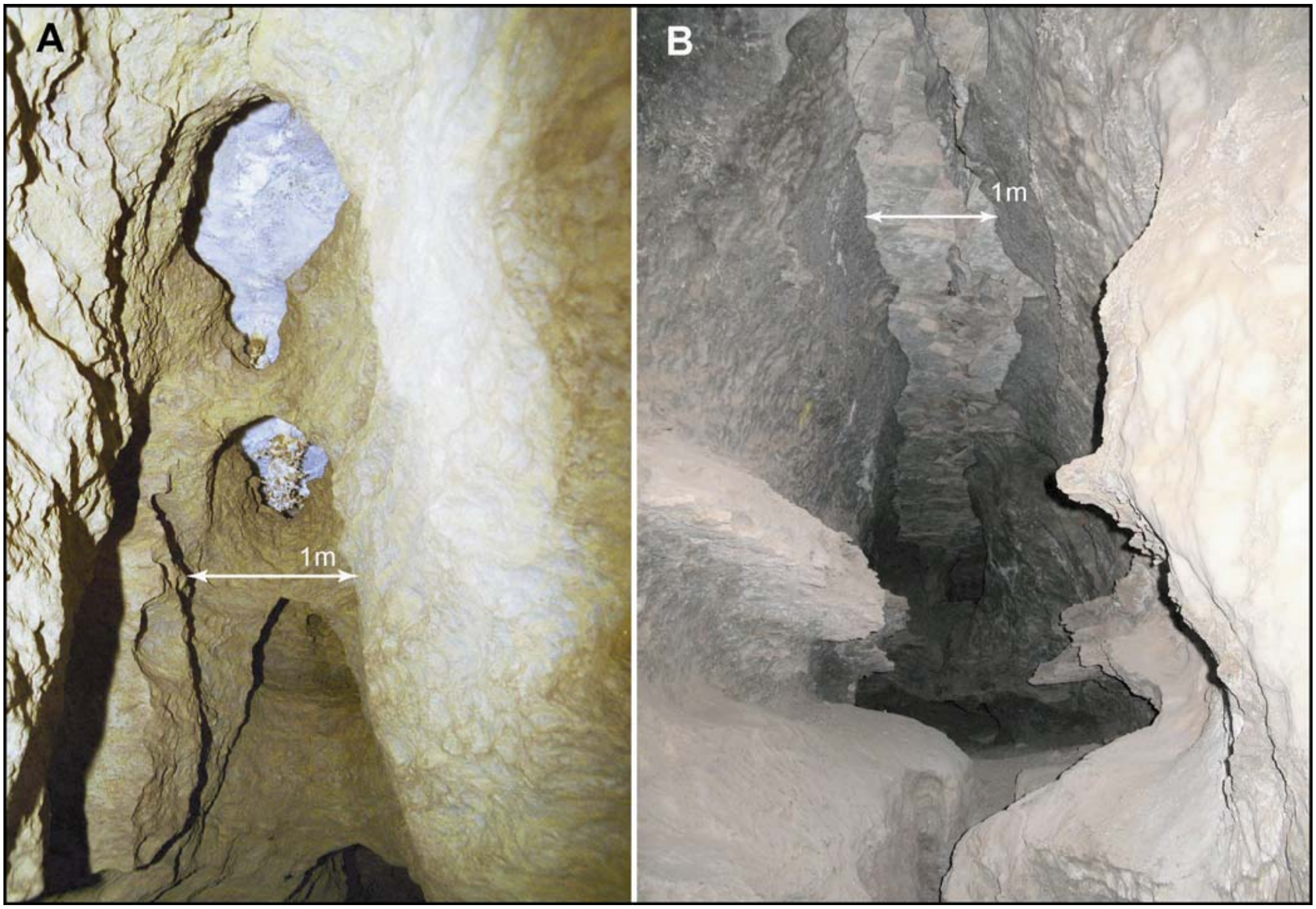

Plate 10. Outlets breaching to the upper discharge boundary (*e.g.* prominent bedding plane or diffusely permeable bed). A = closely spaced, partly merged serial outlet cupolas in the passage ceiling, Fuchslabyrinth Cave, Germany (Triassic limestones); B = the bottom of the upper aquifer, in this case a fractured dolomite bed, exposed at the ceiling along the entire length of a passage, Coffee Cave, Permian gypsum, NM, USA. Note irregular gypsum margins on the underside of the upper dolomite layer in photo B indicating coalescence of serial cupolas**.** Photo by A. Klimchouk.

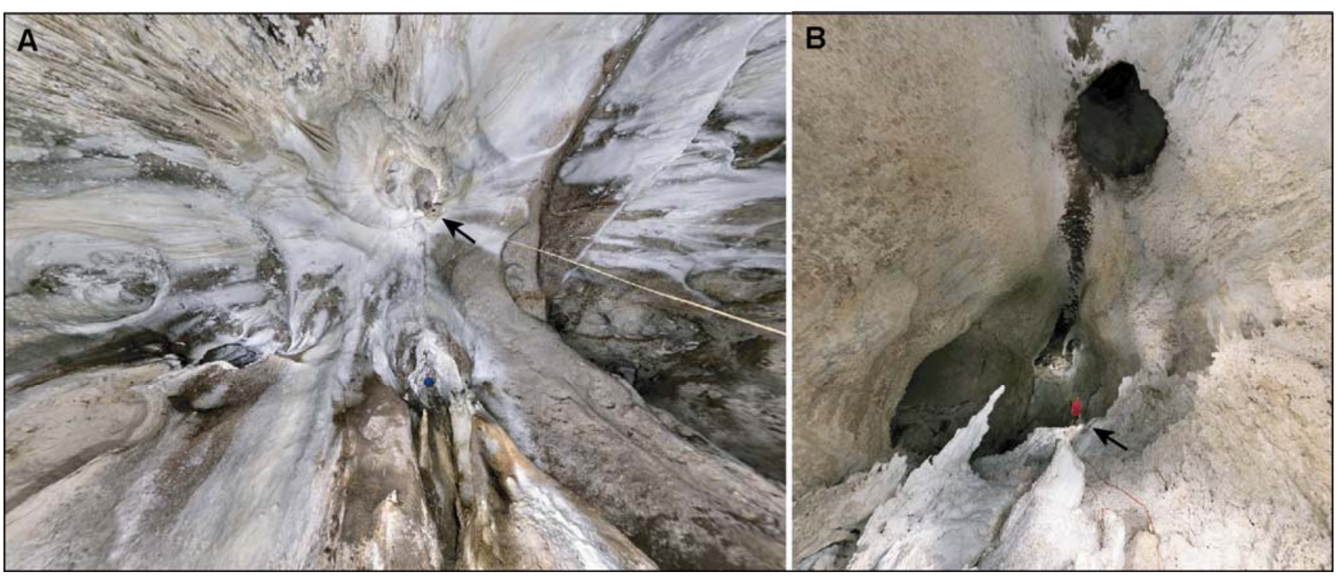

Plate 11. Mega-outlets in Lechuguilla Cave, Guadalupe Mountains, NM, USA (Permian limestone), viewed from below. Arrows point to people for scale. A = Echo Chamber, a mega-outlet at the ceiling (explored by a caver climbing a rope). B = Prickly Ice Cube Room. Note a large ceiling channel rising from a passage to a mega-outlet in photo B. Snapshots from 360^{o} panoramic views of Lechuguilla Cave, NM, by Four Chambers Studio in collaboration with the US National Park Service (with permission).

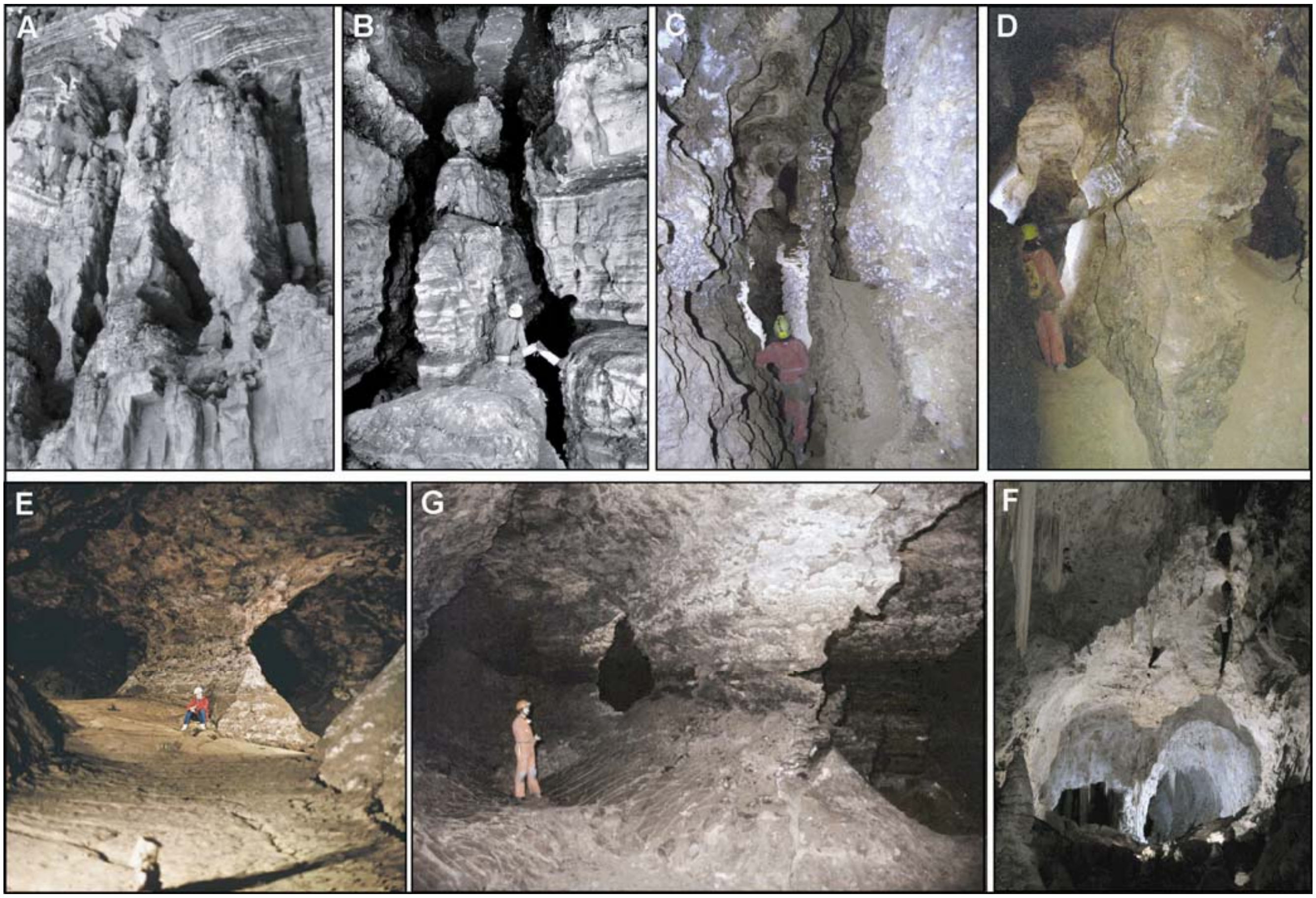

Plate 12. Bedrock partitions between closely-spaced passages in maze caves. A = Parallel slot-like passages opened to the walls of Marble Canyon, Arizona, USA. Top to bottom height is about 65 m (from Huntoon, 2000); B = Slot-like passages with a thin partition in Wind Cave, SD, US (photo by A. Palmer); C and D = Thin partitions between passages in Slavka Cave, western Ukraine (photo by A. Klimchouk); E and G = Thin partitions between passages in Zoloushka Cave, western Ukraine (photos by B. Ridush and V. Kisselev). The pillars were thinned by dissolution at the water table during a late stage of cave development; F = Thin partition between parallel passages in the Lower Cave section of Carlsbad Cavern (photo by A. Klimchouk).

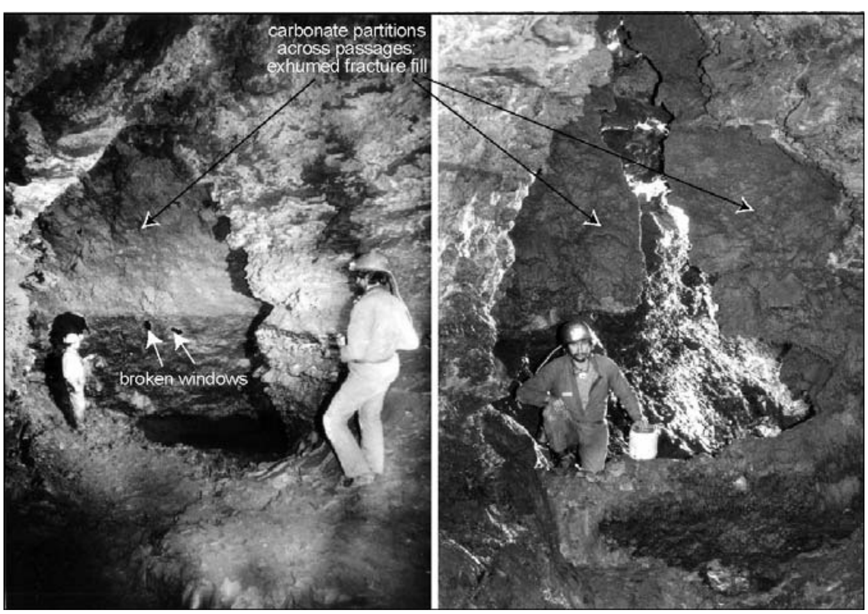

Plate 13. Partitions of carbonate fracture fill, only 1-10 cm thick, completely (left photo) or partially (right photo) partitioning large passages in Zoloushka Cave, western Ukraine. Photos by B. Ridush.

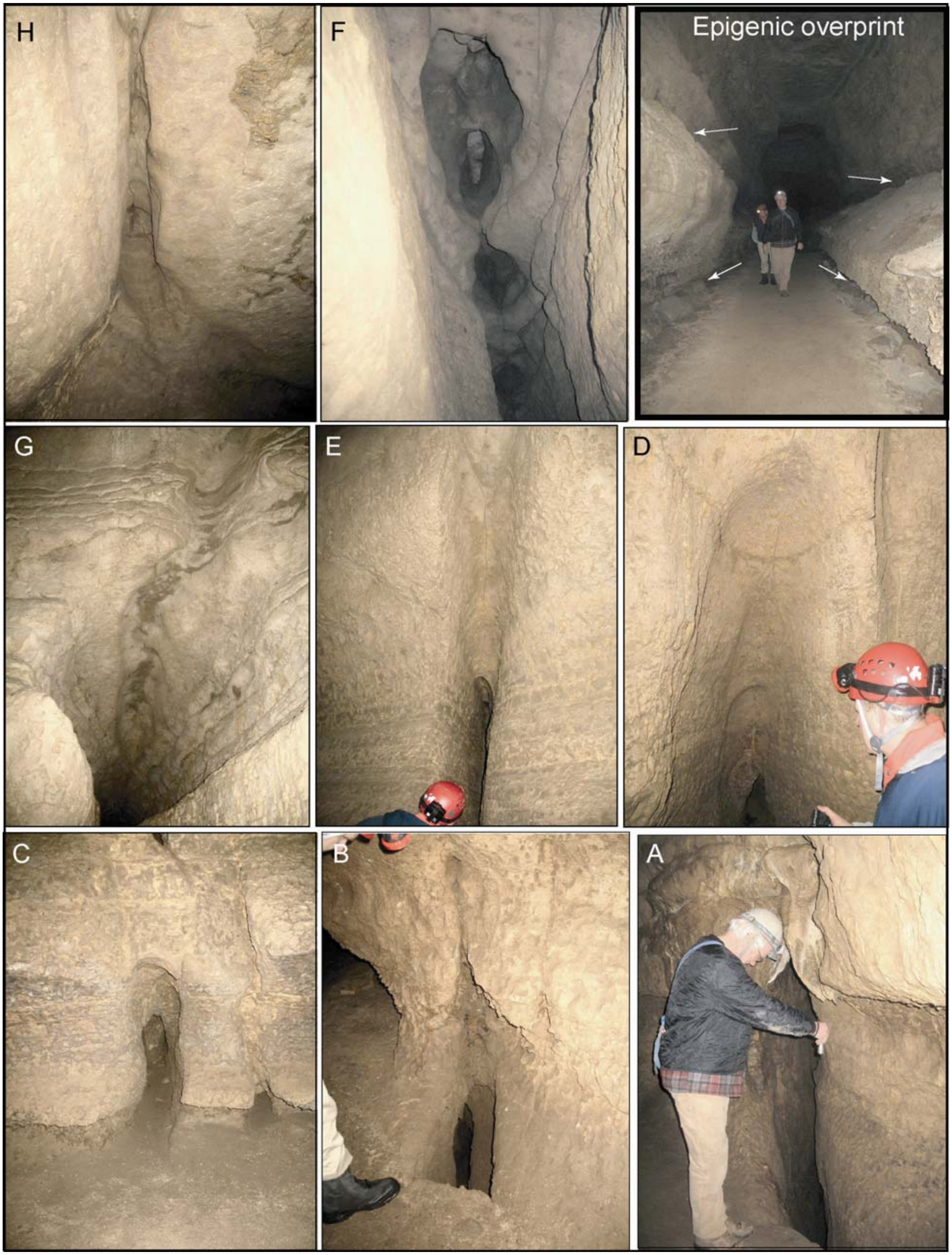

Plate 14. Hypogenic morphology in Mystery Cave, Minnesota. Labelling of photos in the reverse order emphasises functional vertical relationships of the features. A, B and C = feeders; D, E and G = rising wall and ceiling channel; F and H = ceiling channels and outlets; Upper right photo = epigenic overprint. Photos by A. Klimchouk.

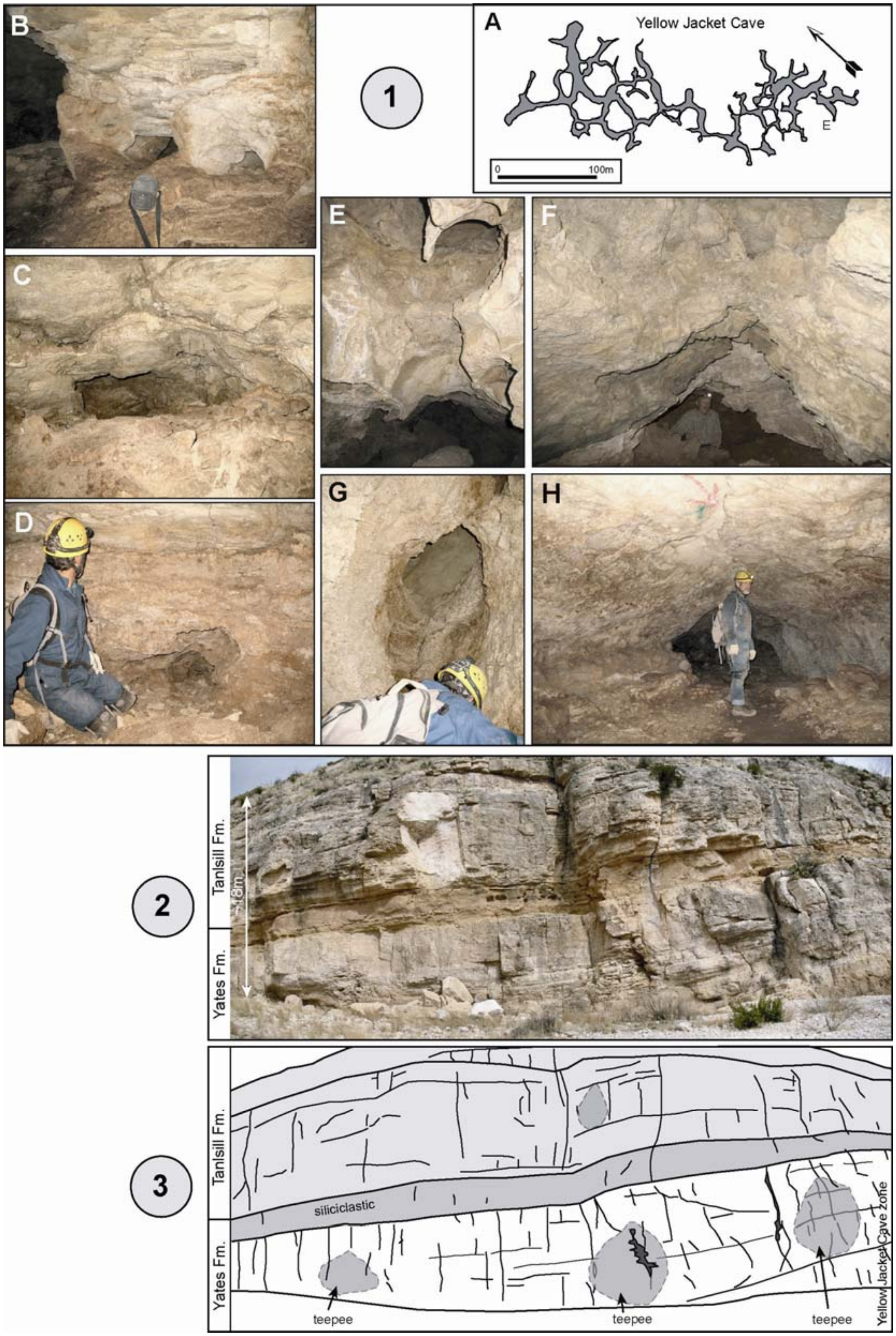

Plate 16. Morphological (1) and geological (2 and 3) characteristics of Yellow Jacket Cave, Dark Canyon, Guadalupe Mountains, NM. The cave is a slightly inclined maze with an unusual polygonal pattern (A = simplified map from the survey by D.Belski and the Pecos Valley Grotto), confined within beds of the upper Yates Formation that contain teepee structures. Passages are controlled mainly by axial fractures of linear teepee structures (photos F and H). The upper limit of the cave is the poorly-permeable siliciclastic bed (photo 2 and line drawing 3 of exposure in proximity to the cave). Note that most of the fractures do not cross the borders of particular beds. Photos B, C and D show feeders, which are numerous throughout the cave. Photos E and G show ceiling channels and outlet cupolas. Photos by A.Klimchouk.

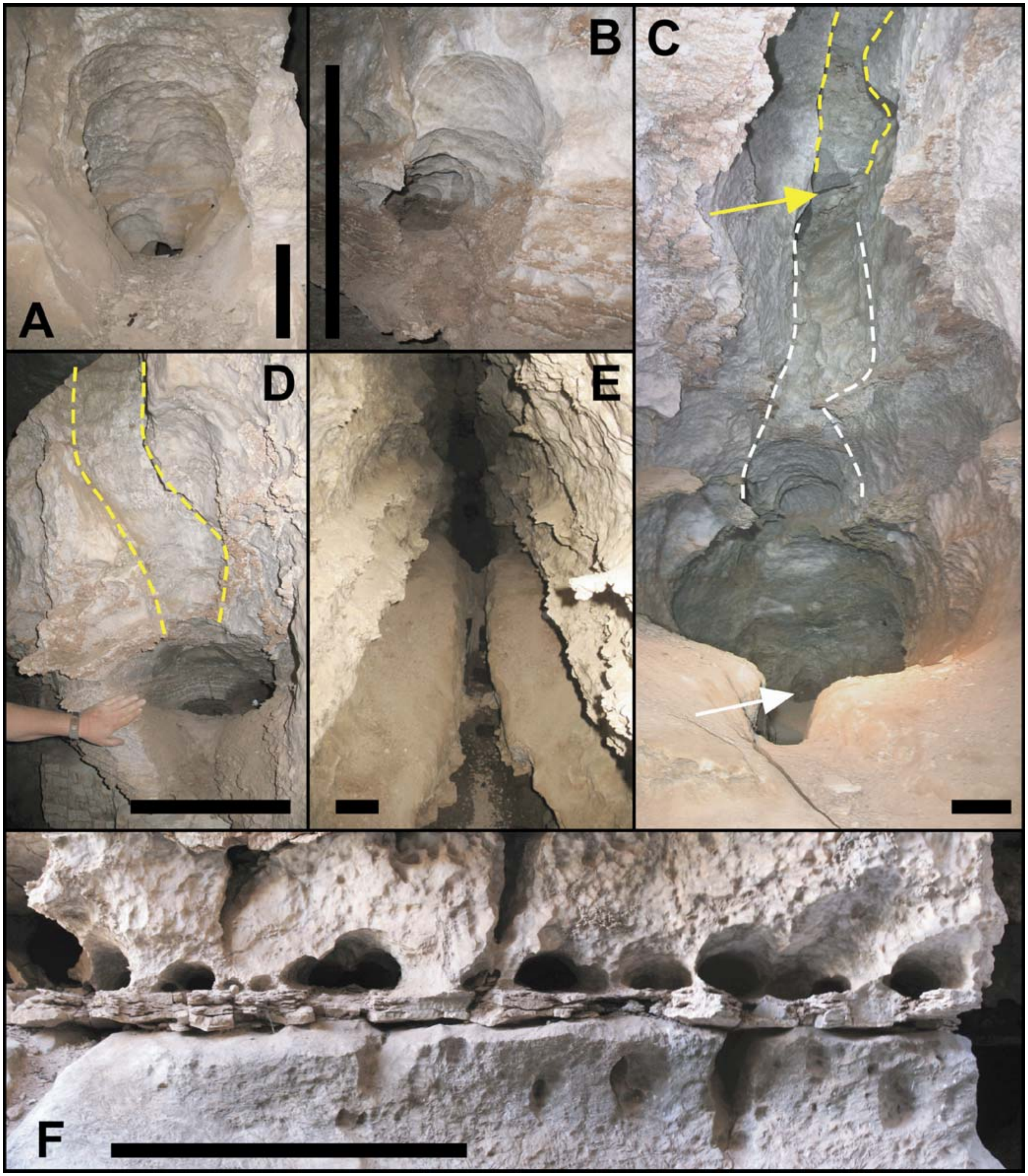

Plate 17. Hypogenic morphology in Coffee Cave, NM, USA (Permian gypsum): Feeder features (photos and compilation by K.Stafford, from Stafford et al., 2008). Black scale bars in figures are all approximately 0.5 m and camera angle is near-horizontal in all feeder features photos. A = point source feeder showing prominent doming morphology proximal to master passage; B = typical feeder showing development at the top of a dolomitic interbed; C = complete hypogenic morphologic suite showing riser (white arrow), wall channel (dashed white lines), ceiling channel (solid yellow lines) and outlet (yellow arrow); D = well developed wall riser with associated wall channel (dashed yellow lines); E = linear riser developed along axis of master passage; F) dense cluster of small feeders above dolomite interbed with minor vadose overprinting below dolomite interbed.

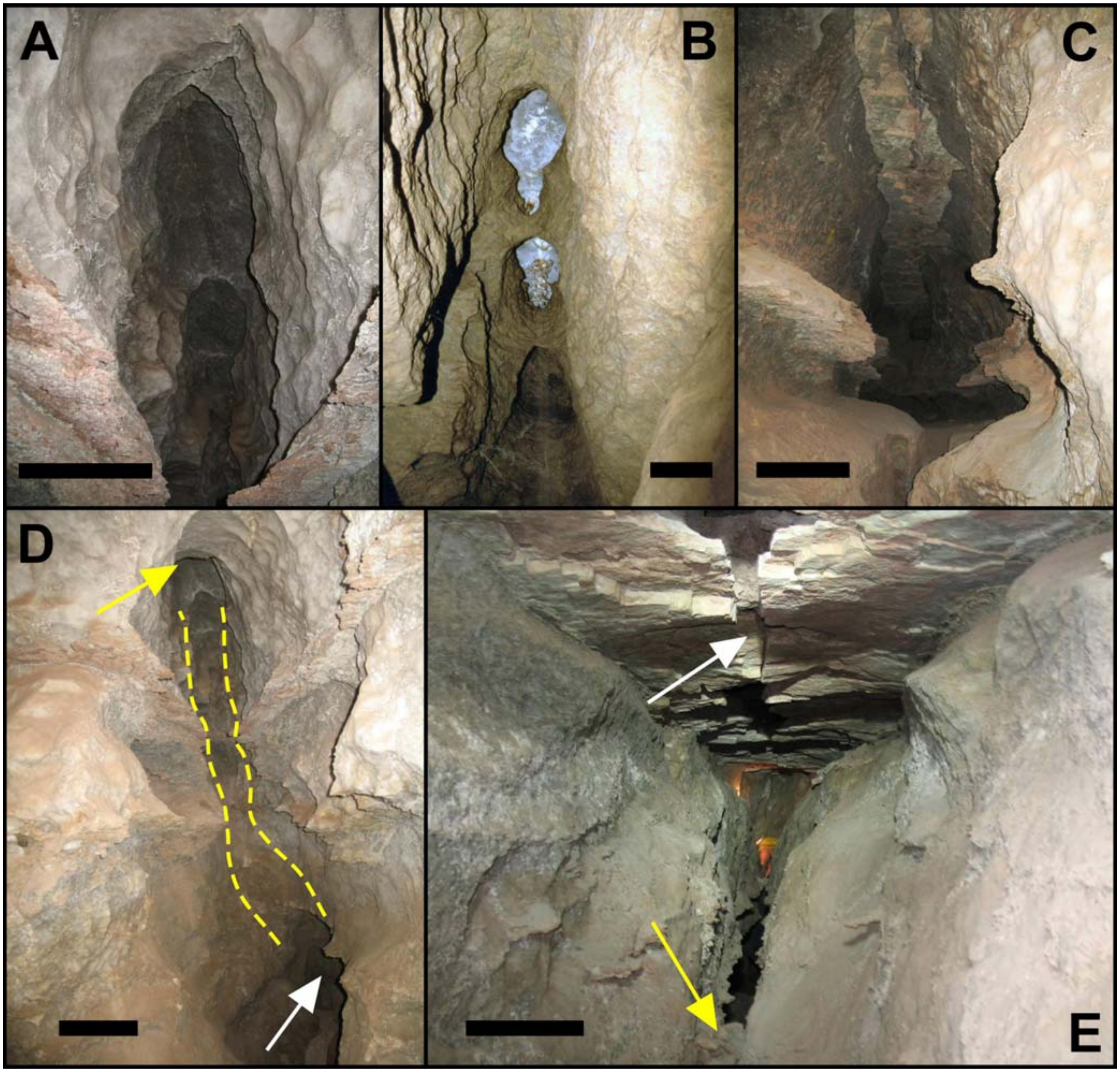

Plate 18. Hypogenic morphology in Coffee Cave, NM, USA: Outlet features (photos and compilation by K.Stafford, from Stafford et al., 2008). Black scale bars in figures are all approximately 0.5 m. A = series of typical ceiling cupolas (camera angle is ~60° up from horizontal, looking toward the ceiling); B = series of cupolas that are in the process of coalescing (camera angle is ~70° up from horizontal, looking toward the ceiling); C = ceiling channel formed by complete coalescing of serial cupolas (camera angle is ~30° up from horizontal, looking toward the ceiling); D = complete hypogenic morphologic suite showing riser (white arrow), wall channel (yellow dashed lines), and ceiling cupola (yellow arrow) (camera angle is roughly horizontal); E = rift-like passage showing linear feeder (yellow arrow), triangular passage and upper dolomite bed that has partially collapsed due to loss of buoyant support (white arrow) (camera angle is roughly horizontal). Photo B (by A. Klimchouk) is from Fuchslabyrinth Cave, Baden-Würtenberg, Germany, instead of Coffee Cave, in order to well-illustrate the intermediate stage of cupola coalescence.

Plate 19. Megasinkholes (shafts) associated with hydrothermal systems, with travertine deposition near water table.
Upper photo: El Zacatón, Mexico, a 325-m deep shaft, approximately 80 m wide at the surface, (see Chapter 4.5; photo by M.Gary).
Lower photo: Kizoren obruk (*obruck* is a generic name for large collapse features in Turkey), a 125-m deep, approximately 120-m wide sinkhole in Konya Basin, Turkey. Travertine is exposed on the walls as the water table declined for ~ 20 m during the last several decades due to regional groundwater extraction. Photo by A.Klimchouk.

4.5 Selected examples of caves formed by hypogenic transverse speleogenesis

In this section, a brief overview is provided of some exemplary caves for which hypogene transverse origin is firmly established or suspected based on available publications and personal observations by the author. Not all of them have been previously interpreted in this way, but their morphological and geological characteristics, consistent with the criteria discussed above, strongly suggest their development by rising flow in confined settings.

In no way should this overview be considered as exhaustive for hypogenic caves. The number of such caves recognized around the world is rapidly growing, and it is going to expand dramatically by re-interpretation of many caves, based on the new approach suggested in this book. Rather, this is a list of reference cases to use in relating, refining or revising the origin of a great number of caves in many regions, and an illustration of the variety of geographic and geologic settings in which ascending hypogenic speleogenesis occurs.

Central and Western Europe

Hypogenic transverse caves are abundant throughout Central and Western Europe, occurring in both cratonic and disrupted (folded) settings.

Probably the most well known and early recognized as hypogenic (hydrothermal) are caves in the Buda Mountains, Hungary (*e.g.* Müller and Sarvary, 1977. There is abundant literature on Hungarian hydrothermal caves, summarized in Takács-Bolner and Kraus (1989) and Dublyansky, Yu. (1995, 2000c). Triassic cherty limestones are overlain by 40-60 m of Eocene limestones covered by marls that form a largely impermeable cap. Denudation and fluvial erosion imposed over the complex block-fault geologic structure created various situations of breaching and draining of the cave-hosting limestones. Mean temperatures of descending waters invading the caves are 8-13^{o} C. Different thermal springs have mean temperatures between 20-60^{o} C, indicating that differing amounts of mixing take place (Ford and Williams, 2007). Almost the full range of thermal cave types and evolutionary stages of development can be found in Budapest, including 2-D and 3-D mazes, relict shafts, chambers, and modern caves discharging rising thermal waters at the level of the Danube River.

Ford and Williams (2007) refer to a series of illustrative examples of the Buda Mountains caves. Pálvölgyi (Figure 22, A) and Ferenc-hegy caves are multi-story mazes. Szemlöhegy Cave is a more simple outlet rift from along joints; it has abundant subaqueous calcite crusts. Molnár János is a modern outlet; its warm waters have been dove to -70 m, discovering ~4 km of maze at depth. Jószefhegy Cave is a shaft descending to ramiform chambers with abundant secondary gypsum deposits, where Ford and Williams (2007) point out that CO_2 (hydrothermal) and H_2S dissolution mechanisms may mingle within a small geographical area.

Sátorkö-puszta Cave and Bátori Cave (Figure 22, B) are examples of basal chambers with a tree-like branching pattern of rising passages containing sequences of spherical cupolas. These passages extend for about 40 m above the chamber in Sátorkö-puszta. The origin of cupolas had been attributed earlier to the convectional condensation mechanism operating above the water table (Szunyogh, 1989). The chamber walls are largely converted to gypsum. Ford and Williams (2007) believe these caves are monogenetic H_2S systems, major enlargements by H_2S processes of earlier rising water shafts.

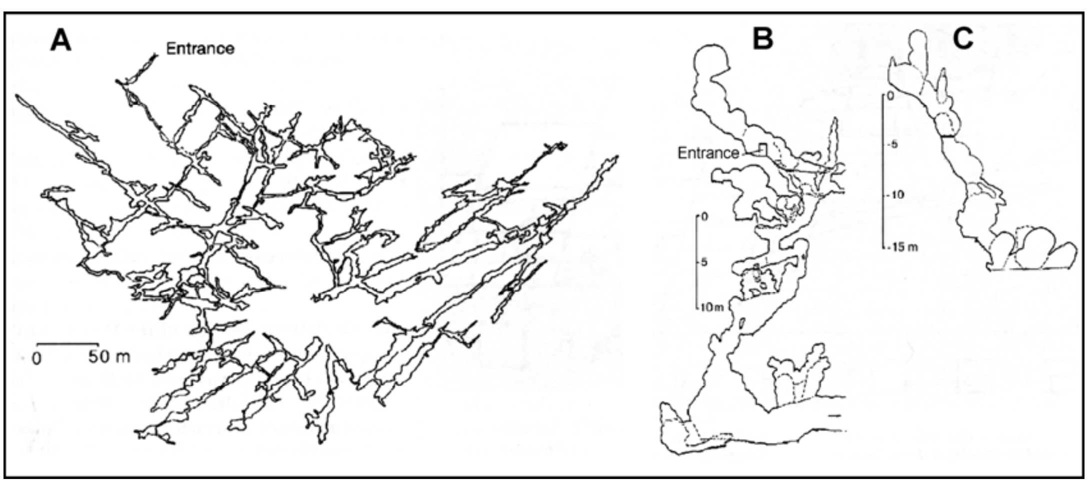

Figure 22. Patterns of hydrothermal caves of the Buda Mountains, Hungary. A = maze pattern, plan view of Pálvölgyi Cave (by J. Kárpát and K. Takácsne-Bolner); B = bush-like patterns, 1 = profile of Sátorkö-puszta Cave (by M. Juhasz), 2 = profile of Bátori Cave (by P. Borka and J. Kárpát). From Dublyansky Yu. (2000c).

Evidence for hydrothermal hydrogeologic and karst features is abundant from carbonate-composed folded structures in many regions of Romania, particularly in the Codru Moma and Bihor Mountains. In the former region, thermal waters rise through Mesozoic limestones, along thrust-related faults and basin-limiting faults, from the underlying dolomites and quartzitic sandstones, producing thermal springs (Oraseanu and Mather, 2000). In the Bihor Mountains, Onac (2002) described skarn-hosted and classical hydrothermal caves, passages, and rooms with specific mineralization encountered by ore mines. Individual caves vary in size and total up to 500 m in length. The skarn-hosted caves are supposed to form during contact metamorphism of limestones, which involves decarbonation and release of CO_2. Later on, caves were formed within the overlying limestones by rising thermal flow, either by retrograde solubility along cooling paths or by mixing of ascending hydrothermal fluids with oxygenated waters of a shallow flow system.

The famous Movile Cave in the Miocene oolitic limestones of the Dobrogea Plateau, the Black Sea area, is assumed to have an epigenic origin related to the drop in sea level of the Black Sea during middle Pleistocene (Lascu, 2004), but the presence of a confined aquifer in the deeper sections of the carbonate sequence with low-grade thermal sulfidic waters, and a highly specific assemblage of cave fauna indicative of an isolated cave environment, suggest the possibility of a hypogenic origin for this cave. This is also supported by the presence of a 4 km long maze cave nearby.

One of the most outstanding examples of deep-seated hypogene speleogenesis is a giant cavity encountered by boreholes in Precambrian marbles in the Rodopy Mountains, Bulgaria (Dubljansky V., 2000). The top of the cavity is intercepted at 560-800 m below the surface, and the greatest measured vertical dimension of the cavity is more than 1340 m (the borehole did not reach the bottom). The cavity has an estimated volume of 237.6 million m^3 and is filled with thermal waters. The highest measured water temperature was 129.6^o C at a depth of -1279 m (359 m below sea level). Hydraulic head is several hundred meters higher than the top of the cavity. Many smaller cavities of hydrothermal origin are known in this area.

Outstanding hypogenic caves, formed by rising thermal H_2S waters, are known in central and south Italy (Umbria and Marche regions), in Jurassic and Paleogene carbonates of the Neogene fold-and-thrust belt (Galdenzi and Menichetti, 1995). Jurassic carbonates are variable in initial porosity, with well-developed sedimentary facies in 4-5 m thick cyclothemic sequences. Some hypogenic caves occur in Pleistocene travertine. They are large 3-dimensional maze systems, in which patterns of basal input zones and points are recognized through fractures and vuggy porosity networks at the bottom of the carbonate sequence where mineralized water rises up from the Triassic evaporitic beds. Their morphology displays characteristic features of uprising flow, including rising shafts, wall and ceiling half-tubes, roof pendants, cupolas and blind domepits, through network and spongework passages situated at different levels. These caves contain massive gypsum deposits, like those found in the caves of the Guadalupe Mountains, USA. Galdenzi and Menichetti (1995) noted the remarkable similarity of internal morphologies in the caves that have different patterns and local geological contexts. Relict hypogenic caves are represented by Monte Cucco caves (31.3 km long and 930 m deep 3-D maze system; Figure 24) and Pozzi della Piana (3.5 km long network maze in two levels within a vertical range of 25 m); whereas sections in some caves, such as those in Frasassi Gorge and Acquasanta Terme, are still active, presently at the water table, with H_2S waters rising from depth. The principal chemical mechanism for hypogenic speleogenesis in the region is sulfuric acid dissolution due to mixing of deep and shallow flow systems.

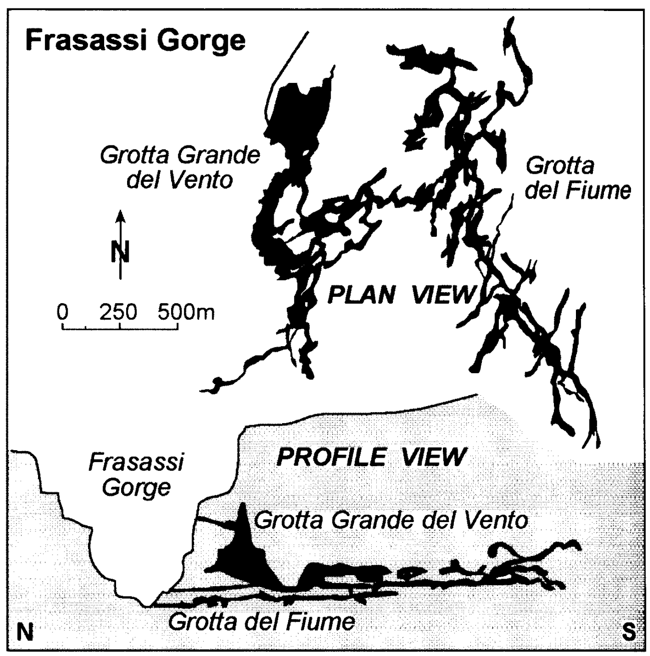

Figure 23. Plan view and profile of Frasassi caves. Note inclined cave stories in the profile (from Hose and Macalady, 2006).

Some caves that display quasi-horizontal levels (*e.g.* the Frasassi caves; Figure 23) were interpreted to form in the shallow phreatic zone near the water table, where H_2S oxidation is most intense in the present setting. However, Galdenzi and Menichetti (1995) have noted difficulties in explaining the upper levels of the Frasassi caves by analogy with the present water table/shallow phreatic conditions. Many other caves have spatial patterns that do not show any apparent relation to the water table. The vertical structure of Acquasanta Cave and other caves in the Rio Garrafo Gorge (Figure 25), entirely developed under a low-permeability cover, seems to be controlled by lithostratigraphy and faulting in the core of a formerly confined but now breached anticline, showing convergence of the cave development to the breaching point. The caves of Monte Cucco (Figure 24) demonstrate a largely structurally controlled pattern. Together with clearly phreatic (hypogenic) morphology and stratigraphically controlled vertical heterogeneities in permeabilities, this suggests a somewhat confined cave-forming environment. The Faggeto Tondo cave lies under a low-permeability level of nodular limestone and developed concordant to bedding within a vertical range of about 300 m. La Grotta shows the most complex 3-D structure, with multiple inclined levels within a vertical range of about 930 m. The Parano Gorge caves are developed largely beneath very thick, low-permeability marls and sandstones of Miocene age, partly breached by a gorge (Figure 26). The cave system has a complex 3-D pattern with multiple levels, controlled by stratigraphy and fractures (Galdenzi and Menichetti, 1995).

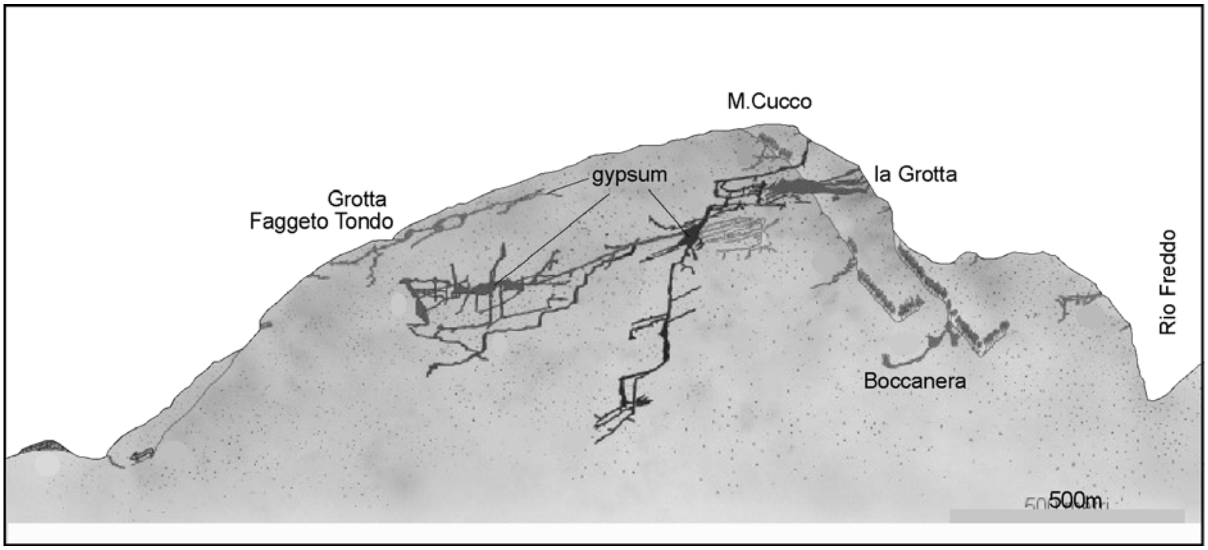

Figure 24. Schematic cross-section through the Monte Cucco karst system, central Italy (courtesy of Centro Excusionistico Naturalistico Speleologico Costacciaro: www.sens.it). The main occurrences of massive gypsum are indicated after Galdenzi and Menichetti (1995).

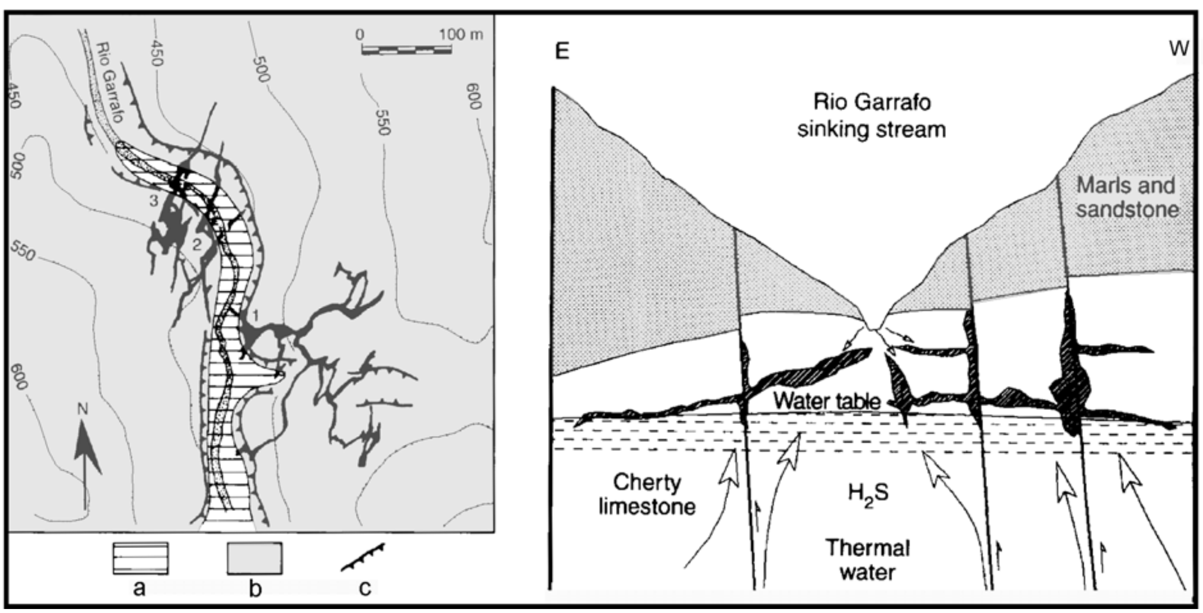

Figure 25. Speleological map on left and schematic cross-section of Rio Garrafo Gorge on right, showing occurrence of Acquasanta caves. Key: a = marly, cherty limestone; b = low-permeability cover (marls and sandstone). From Galdenzi and Menichetti (1995).

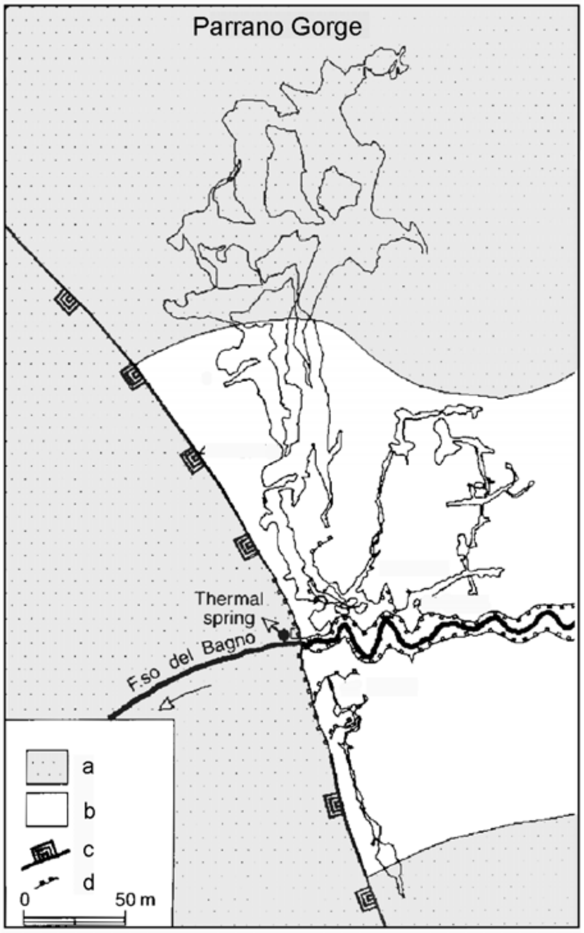

Figure 26. Speleological map of Parrano Gorge, Central Italy. Key: a = low-permeability cover of marl and sandstone; b = cherty limestone; c = normal fault; d = morphological scarps (from Galdenzi and Menichetti, 1995).

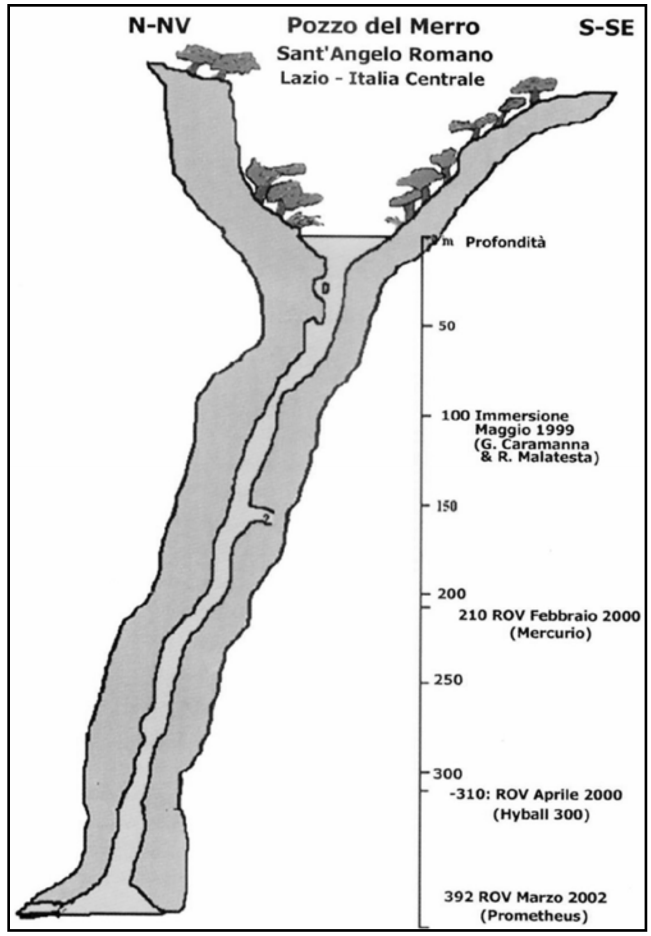

Figure 27. Phreatic shaft, Pozzo del Merro, Italy (from Caramana, 2002).

Pozzo del Merro near Rome, Italy, is the deepest underwater shaft, presumably formed by rising thermal CO_2 and H_2S water. It has been explored by ROV to -392 m underwater and has a total depth of about 500 m, including the entrance sinkhole. It shows the morphology of a rising shaft (Figure 27), in contrast with the roughly cylindrical morphology of Sistema El Zacatón sinkholes in Mexico (see below) where hydrothermal cavities at depth are assumed to open to the surface through collapse.

In south Italy, in northern Calabria, hypogenic caves formed by rising flow are known in isolated limestone massifs (Triassic through Cretaceous) surrounded by clastic Pliocene sediments, in the vicinity of the Sangineto transform fault, a collision zone between the European and African plates (Galdenzi, 1997). The caves have 3-D structures, rising morphology and abundant gypsum deposits. Some caves reach the current water table, with sulfur-rich water at 40^{o} C. Despite their chemistry and high temperature, isotopic signatures of the deep waters demonstrate a meteoric origin.

Hypogenic caves of chamber and maze types have been recently reported in strongly folded Alpine Jurassic limestones from Provence, France (Audra *et al.*, 2002). The Adaouste (a 3-D maze) and Champignons (a chamber with deep rifts) caves are shown to have formed by rising thermal artesian waters. They display the key elements of the morphologic suite of rising flow and specific hydrothermal secondary formations. Complex evolution is reconstructed for these caves, with major periods of hypogene speleogenesis assumed to be Miocene (older than 11 Ma) for Champignons and Upper Tortonian (8.5 to 5.8 Ma) for Adaouste. Chevalley Aven and Serpents Cave are active thermal-sulfidic caves presently at the water table stage in the Bauges massif in the northern French Pre-Alps (Audra *et al.*, 2007b). They have convection spherical cupolas at the ceiling, active condensation-corrosion processes, deposition of calcite rims and replacement gypsum and popcorn. Audra *et al.* (2007b) leave an open question as to whether the ceiling pattern is created by condensation-corrosion in the vadose zone or if condensation-corrosion is occurring in older conduits of a phreatic origin.

In their recent review of the Alpine karst in Europe, Audra *et al.* (2007a) point out that hypogenic (hydrothermal) karst appears more widespread in the region than previously assumed. Rising hydrothermal systems are usually located near thrust and strike-slip faults. The hydrothermal origin of these caves can be recognized from their characteristic convection-related morphology and presence of calcite spar. Ascending hydrothermal systems create well-connected cave systems, which later are generally reused and reshaped by epigenetic speleogenesis, overwriting many marks of their hypogenic origin. These were conserved, however, when caves were rapidly fossilized (Audra *et al.*, 2007a).

Instructive examples of densely packed hypogenic mazes in Western Europe are Fuchslabyrinth Cave (6.4 km; Figure 28-A; Müller et al, 1994) in Baden-Würtenberg, Germany, and Moestroff Cave (4 km; Figure 28-B; Massen, 1997) in Luxemburg. Both have a main story and lower story. The caves were developed within particular beds of the stratified carbonate Muschelkalk Formation under former artesian conditions. The presence of low permeability beds in the caprock prevents vertical downward percolation to the caves, which perfectly display the morphological suite of rising flow, described in Section 4.4 (see Plate 1-C, Plate 3-H, Plate 10-A). No mineralogical evidence is found pointing to the involvement of H_2S waters, but dissolution due to mixing or rising flow and lateral flow is clearly an option. Largely similar network mazes are known in the Carboniferous limestones of the Northern Pennines, United Kingdom, some of them intercepted by the denudation surface while others have been encountered by mines (Ryder, 1975). The multi-story Knock Fell Caverns (4 km; Figure 28-C) has a substantial overprint of water table and vadose features but the morphologic suite of rising flow is still easily recognizable.

Among gypsum caves in Western Europe, characteristic examples of hypogenic transverse speleogenesis are Estremera Cave in the Neogene gypsum of the Madrid Basin, Spain (Almendros and Anton Burgos, 1983), and Denis Parisis Cave in the Tertiary gypsum of the Paris Basin, France (Beluche et al, 1996). In Estremera Cave (3.5 km; Figure 29, right) the morphologic suite of rising flow is firmly identified. Denis Parisis Cave (3.5 km; Figure 29, left), a joint-controlled stratiform cave, is encountered by a gypsum mine, being totally isolated from both lateral and downward potential recharge sources. Based on published photographs, the characteristic morphology of "ascending" artesian mazes is recognizable.

A type location for hypogenic speleogenesis in gypsum with cave development at the base of a thick soluble formation by buoyancy-driven dissolution from the basal aquifer, is the South Harz in Germany where large isolated chambers and other irregularly-shaped cavities (Schlotten) were encountered by mines at depths up to 400 m (Kempe, 1996; Figure 30).

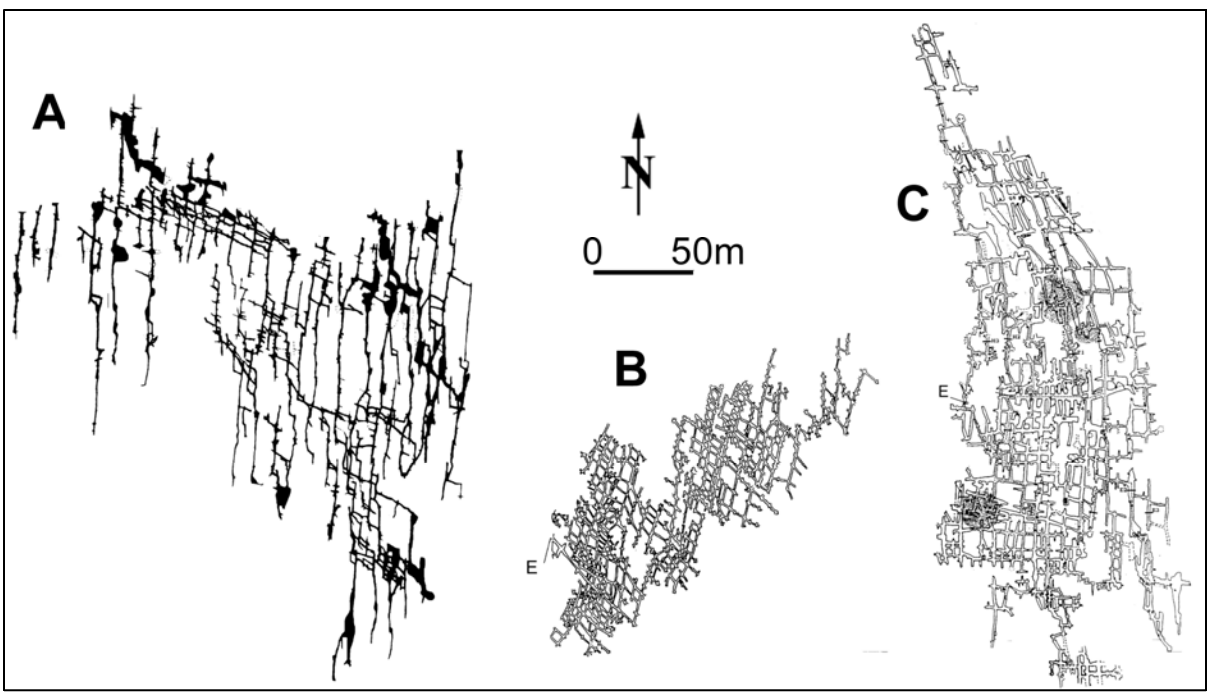

Figure 28. Hypogenic maze caves in Cretaceous (A) and Triassic (B and C) limestones in Western Europe. A = Fuchslabyrinth Cave, Germany (6.4 km; from Müller et al, 1994); B = Moestrof Cave, Luxembourg (4 km; from Massen, 1997); C = Knock Fell Cavern, UK (4 km; from Elliot, 1994).

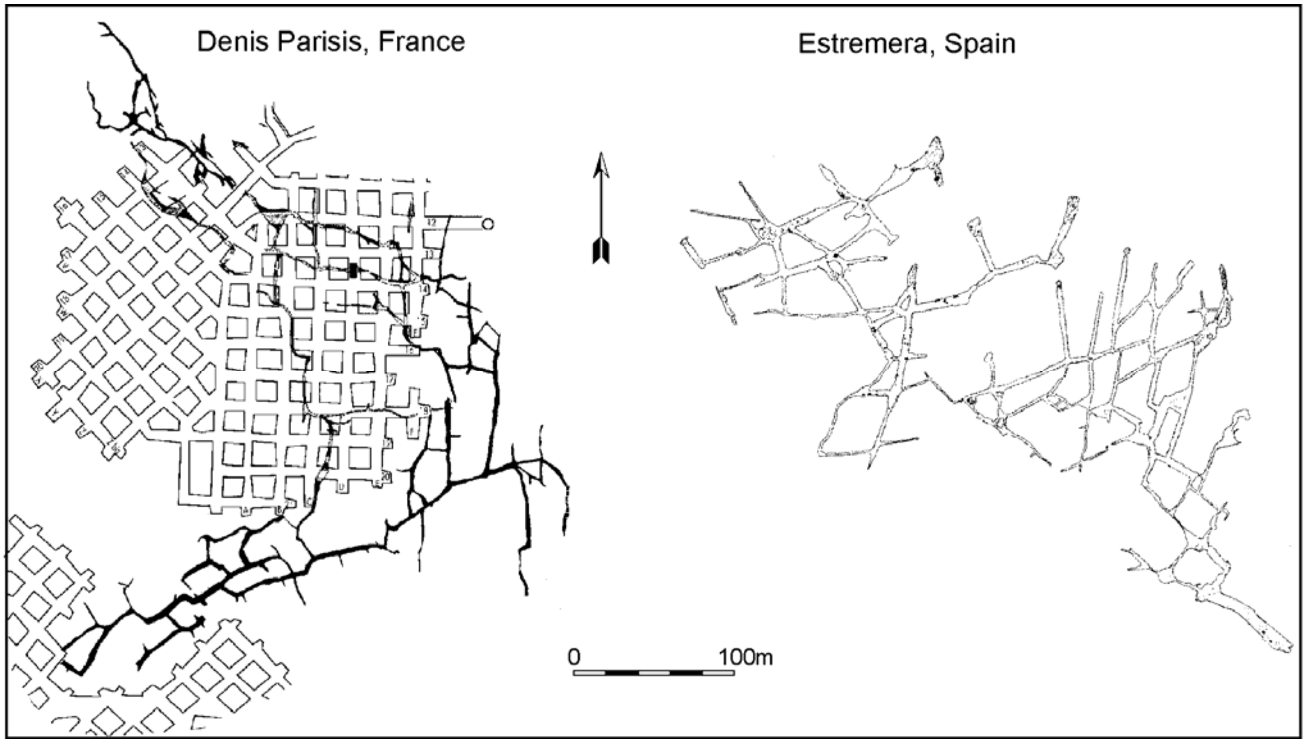

Figure 29. Hypogenic confined mazes in gypsum in Western Europe. Denis Parisis Cave, Paris Basin, France, is a 3.5 km long cave encountered by mines in the Tertiary gypsum (from Beluche et al, 1996). Estremera Cave, Madrid Basin, Spain, also has a length of 3.5 km and occurs in the Neogene gypsum (Almendros and Anton Burgos, 1983).

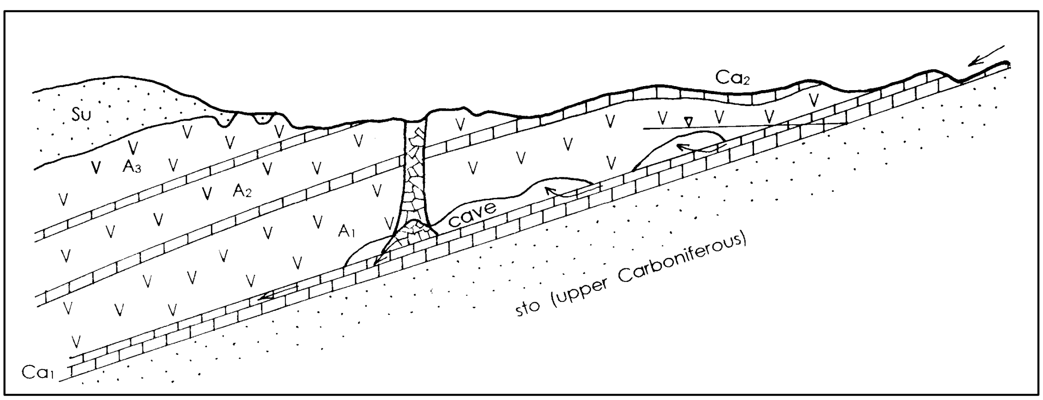

Figure 30. Development of hypogenic caves at the base of a sulfate formation due to buoyancy-driven dissolution, example from South Harz, Germany (from Kempe, 1996).

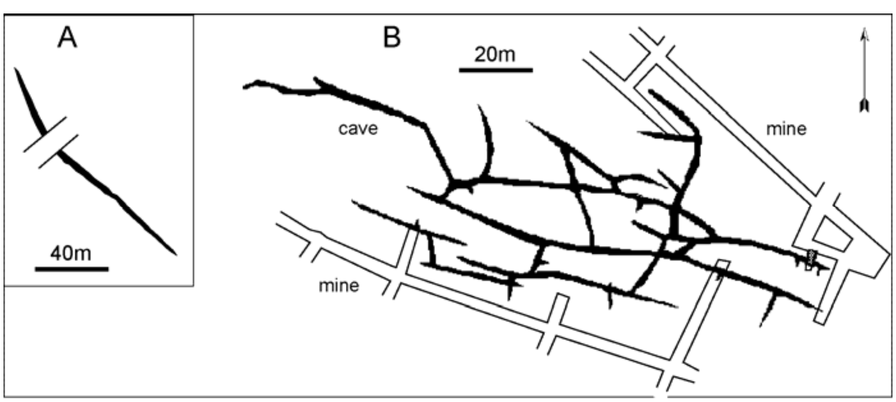

Figure 31. Two-dimensional caves formed by transverse flow across a single bed of Miocene limestone, Prichernomorsky artesian basin, south Ukraine (from Pronin, 1995). Numerous slot-like single-passage caves (A) and small network clusters (B) are encountered by extensive limestone mines beneath Odessa city. Passages are terminated laterally by pinching-out of fissures.

Eastern Europe

The great artesian basins of the Eastern European craton provide many excellent examples of hypogene transverse speleogenesis.

In the Prichernomorsky artesian basin, south Ukraine, beneath the city of Odessa many small caves were intersected by extensive mines in a single limestone bed within the Miocene carbonate sequence, a drained part of the regionally extensive artesian system. The caves represent isolated slot-like passages (Figure 31-A) and crude clusters of intersecting passages (B), the longest cave being a relatively small maze with 1.4 km of mapped passages. These caves are simple and unambiguous examples of transverse speleogenesis; most passages laterally terminate as narrow, apparently declining fissures. They were formed by direct flow between the lower and upper boundaries of a particular limestone bed that contains a single-story intrastratal set of fractures, which are poorly connected laterally.

Among the world's foremost examples of hypogene (confined) transverse speleogenesis are the extensive caves in the Miocene gypsum in the western Ukraine (Figure 32). These are 3-D (multiple-story) network mazes confined within a single 16-20 m thick gypsum bed, sandwiched between two aquifers. The study of their patterns and morphology, along with the regional hydrogeologic analysis, have served as a foundation that firmly established the artesian transverse origin for the caves and the conceptual framework presented in this book (Klimchouk and Rogozhnikov, 1982; Klimchouk and Andrejchuk, 1988; Klimchouk, 1992, 1994, 1996c, 2000b). The caves and their regional settings are discussed elsewhere in this book, and more details can be found in the cited sources. It is important to mention, however, that in the confined zone of the same aquifer system, numerous cavities encountered by exploratory drilling show morphometric characteristics and distribution (in both plan view and cross-section) consistent with the patterns of the explored relict caves (Klimchouk, 1997c).

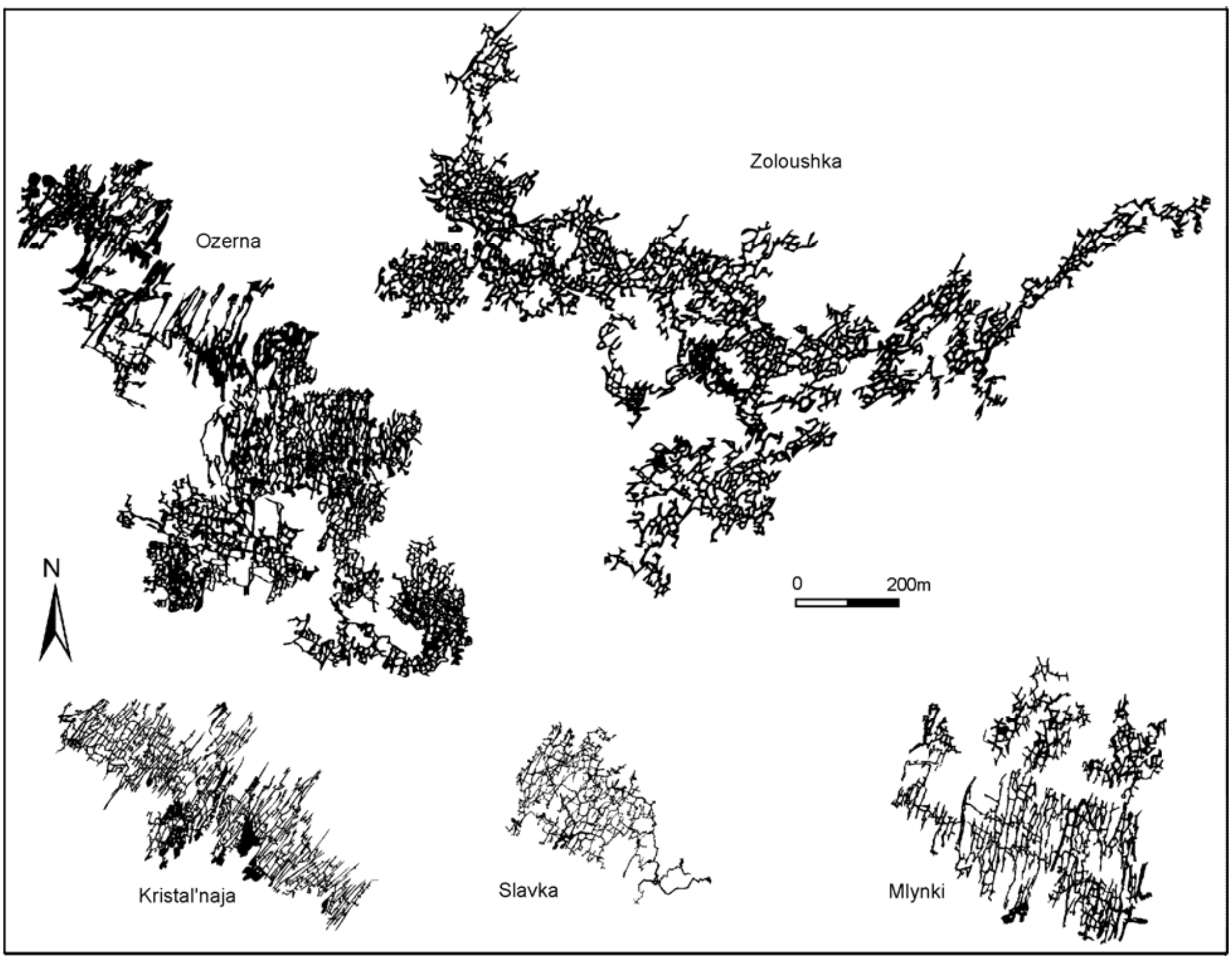

Figure 32. Patterns of hypogenic maze caves in Miocene gypsum in the western Ukraine. The region contains the five longest gypsum caves in the world. The longest is Optymistychna, with 214 km of mapped passages (Figure 12). The second and third longest gypsum caves are Ozerna (117 km) and Zoloushka (90.2 km) shown here. Maps are courtesy of the speleological clubs of Ternopil' (Ozerna, Kristal'naja, and Mlynki), Chernivtsy (Zoloushka), and Kiev (Slavka).

In the eastern outskirts of the Eastern European craton, in the fore-Urals region, Russia, maze caves of hypogene origin are known in both limestones and gypsum. Kungurskaja Gypsum Cave (5.6 km) is a good example of an ascending maze considerably modified with lateral enlargement of passages by backflooding of the nearby Sylva River.

Siberia

In Siberia, a remarkable example is the 60.8 km long two-dimensional network maze of Botovskaya Cave, developed in a Lower Ordovician limestone bed sandwiched between sandstone aquifers (Filippov, 2000; Figure 33 – top map). The area is now an entrenched and drained part of the Angaro-Lensky artesian basin. There are some other maze caves in the region, which likely share the same origin. Hypogene origin is suspected for a number of complex 3-dimensional caves known in other regions of Siberia in limestone (*e.g.* Dolganskaya Yama, 5.12 km long, developed with a vertical range of 125 m in the Riphean - Lower Cambrian limestones of the Vitim Upland) and conglomerates (*e.g.* Bol'shaya Oreshnaya Cave, 47 km long, developed with a vertical range of 250 m in the Ordovician carbonate conglomerates of the Eastern Sayan Upland; Figure 33 – bottom map).

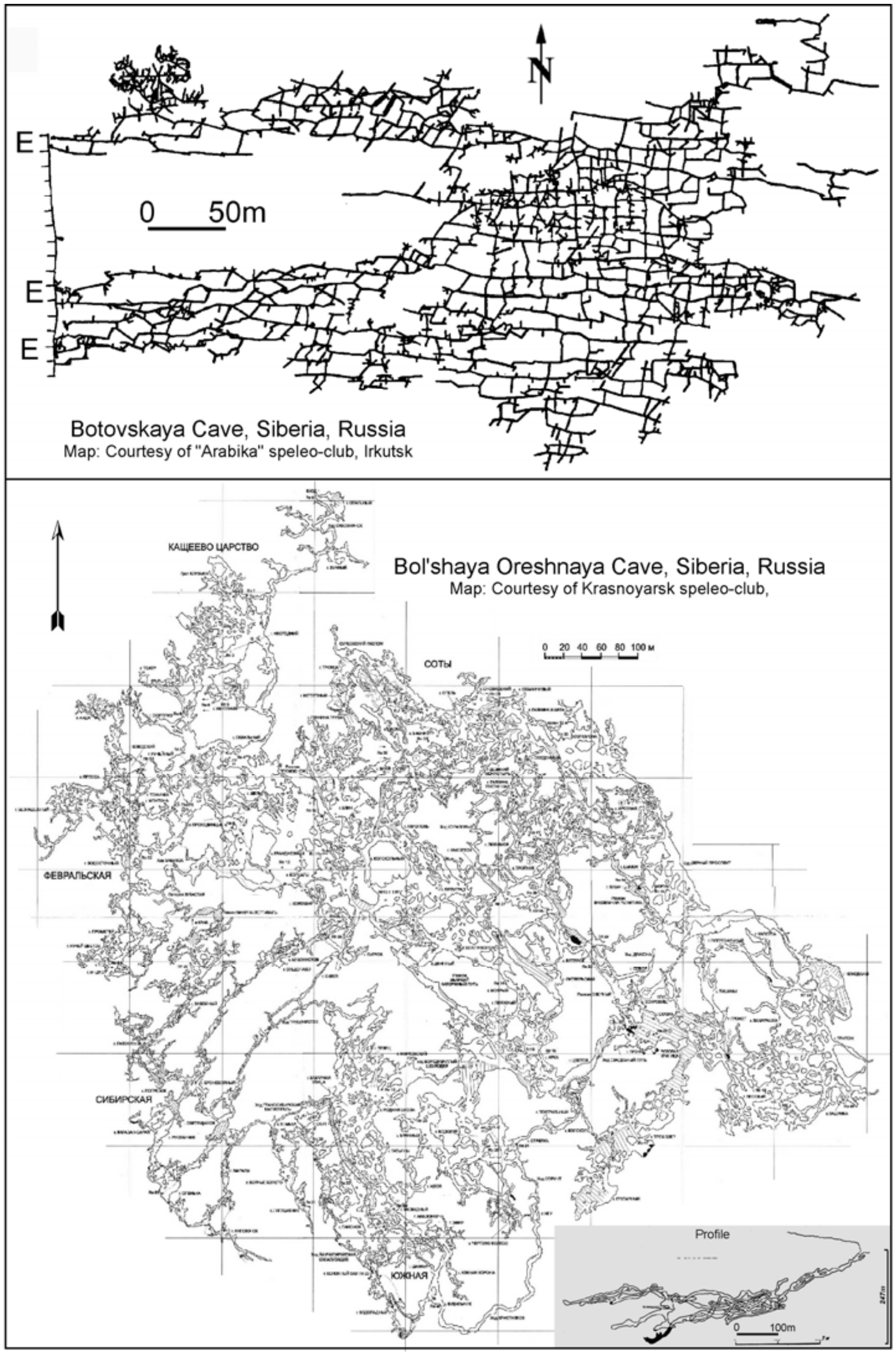

Figure 33. Maze caves in Siberia. Botovskaya Cave (upper map) is a 60.8-km long 2-dimensonal network maze in a Lower Ordovician limestone bed sandwiched between permeable sandstones in a stratified succession of the Angaro-Lensky artesian basin (Filippov, 2000). Bol'shaya Oreshnaya (lower map), Eastern Sayan is a 47 km long cave with a vertical range of 250 m in Ordovician carbonate conglomerates (map courtesy of Krasnoyarsk Speleo-Club).

Caucasus, Central Asia and Middle East

Hypogenic caves are found in both the Great Caucasus and the Minor Caucasus, throughout southwestern Russia, Georgia and Armenia. In the northern ridges of the Great Caucasus some deep sub-vertical caves display odd patterns and hydrothermal mineralization in their deep parts, although these aspects remain poorly studied. In the southern ridges, an outstanding cave of established hydrothermal origin (Dublyansky V., 1980) is Akhali Atoni Cave (Novoafonskaya) in Abkhazia, characterized by the huge volume of its chambers. In the Minor Caucasus, Armenia, an interesting hypogenic cave is Archeri (3.7 km long and 130 m deep), which is an inclined system of wide cavities at five levels along bedding planes and almost entirely lined with a crust of palisade hydrothermal calcite.

Hypogenic caves, mainly produced by hydrothermal and sulfuric acid speleogenesis, are abundant throughout Central Asia (Turkmenistan, Uzbekistan, Tajikistan, and Kyrgystan). Opened by surface denudation, small relict caves are very common throughout the mountains of Tien Shan, Pamir, and Kopetdag. Some large caves are known, both fossil and active, *e.g.* Syjkyrdu Cave in Pamir, Cup-Coutunn/Promezhutochnaja, Geophizicheskaja, Khashim-Ouyuk, and other caves in the Kugitang Range in Tien Shan, and Bakharden Cave in Kopetdag.

The longest cave in Iran, Ghar Alisadr (11.44 km), is developed in the Jurassic Sanandaj-Sirjan Formation. The cave has a joint-controlled maze pattern with passages in several levels and numerous ceiling cupolas (Laumanns et al, 2001). The cave contains numerous lakes representing the current water table and sluggish flow conditions, with lower parts of the cave extending below the water. No specific speleogenetic studies have been done for the cave so far, allowing inference about specific dissolution mechanisms involved. Based on available information about cave pattern and morphology, it is likely that Ghar Alisadr falls into the category of hypogenic transverse speleogenesis.

In Israel, the hypogenic transverse speleogenesis concept has been successfully applied to interpret cave origins at a regional scale (Frumkin and Fischhendler, 2005) and to resolve some important issues of regional hydrogeology (Frumkin and Gvirtzman, 2006). These cited works are particularly instructive as they offer one of the few examples of a regional speleogenetic analysis based on extensive and consistent datasets and performed through a spectrum of paleohydrogeological conditions.

Caves that have no genetic relationships to the surface are estimated to comprise about 95% of the caves in Israel. Many of them are network mazes; others are isolated chambers. Maze caves occur within the massive limestone and dolomite beds in upper sections of the late Cretaceous Judea Group (Bina Formation), beneath the overlying leaky confining Mt. Scopus Group (massive chalk and marls). Numerous maze caves (the longest is 3.45 km long) are distributed along the retreating edges of the confining cover and beneath the cover in the vicinity of deep underground faults and related flexures (Figure 34-A; Frumkin and Fischhendler, 2005). Caves are opened naturally by denudation and artificially by mining and construction holes. The lack of maze caves far away (>0.5 km) from the cover is due to erosional removal of the Bina Formation that contains them. Network patterns are arranged in several stories (up to three), forming 3-D systems. Stories are concordant with bedrock stratification, with horizontal networks in horizontal beds and inclined networks in inclined beds. Maze networks display many laterally blind terminations, abrupt changes in morphology and high passage density (Figure 36).

The study of Frumkin and Gvirtzman (2006) in the western, still largely confined, sector of the same region has shown that hypogene transverse speleogenesis is currently operative in the confined area, being responsible for the “Ayalon Saline Anomaly” (ASA) at the central part of the major Yarkon-Taninim aquifer (Figure 34-B). Analysis of data from quarries and more than 10,000 boreholes suggests locally high porosity within the Bina Formation and shows that stratigraphic distribution of intercepted voids is similar to what is observed in the unconfined area (Figure 35). The analysis of hydrogeologic data indicates that the ASA contains “hot spots,” associated with transverse cave clusters of high permeability, through which warm-brackish groundwater rises to the upper aquifer section (Bina Formation). In these spots, water is substantially warmer and enriched in Cl and H_2S, and has lower pH values as compared with waters a few hundred meters away.

Other than maze caves, isolated chambers are also common for the region, occurring mainly in the more massive carbonates of the Weradim and Kefar Shaul units below the Bina Formation. They range between a few tens to a few hundred meters in their lateral dimensions and usually have a shaft or domepit rising upward from their ceilings, but tapering or terminating at certain higher intervals. Frumkin and Fischhendler (2005) attributed the chambers' origin to mixing dissolution in sluggish phreatic conditions below a water table, where downward vadose percolation locally occurs. This suggestion is based mainly on the uniform occurrence of isolated chambers through the whole area, including areas far away from the confining cover, where currently water table conditions prevail. However, chamber-type cavities also occur in the confined area (Frumkin and Gvirtzman, 2006), and some caves show combined maze-chamber patterns. An alternative model for chamber-type cavities is that they have also formed by rising flow as part of transverse three-dimensional cave systems. Predominant association of mazes and chambers with different hydrostratigraphic units (Figure 35) is consistent with the role of vertical heterogeneities of initial porosity structure, as discussed in Section 3.4. Regular vertical variations in the distribution of elementary cave patterns are a common feature of 3-D hypogenic systems. The differences in the distribution between mazes and chambers within the exposed sections of the Judea Group can be explained by the fact that maze caves are exclusively associated with the Bina Formation, which is removed by denudation at some distance from the cover edge (see Figure 34-A).

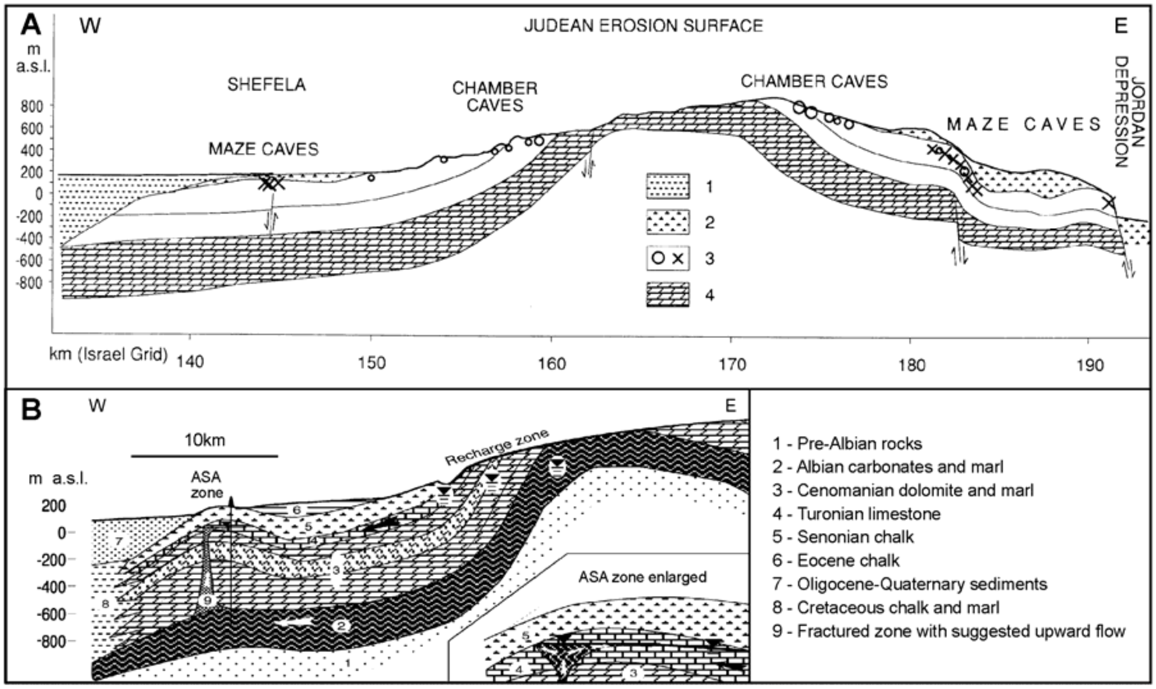

Figure 34. A = Geological section across the Judean arch, Israel. Symbols in boxes: 1 = Cenozoic; 2 = Senonian – early Cenozoic (mostly chalk, confining unit); 3 = dolomite and limestone: Amminadav to Bina Formations of late Cenomanian – Turonian age, with locations of maze caves (X) and chamber caves (O); 4 = Cretaceous dolomite, limestone, and marl, older than Amminadav Formation.
B = Schematic east-west hydrogeological cross-section showing the conceptual model for groundwater flow and hypogenic transverse speleogenesis, the Yarkon-Taninim aquifer and the ASA zone. This cross-section runs roughly parallel to the above section of the Judean arch, but ~ 15 km to south. White arrows = thermal water; black arrow = non-thermal fresh water. Combined from Frumkin and Fischhendler (2005) and Frumkin and Gvirtzman (2006).

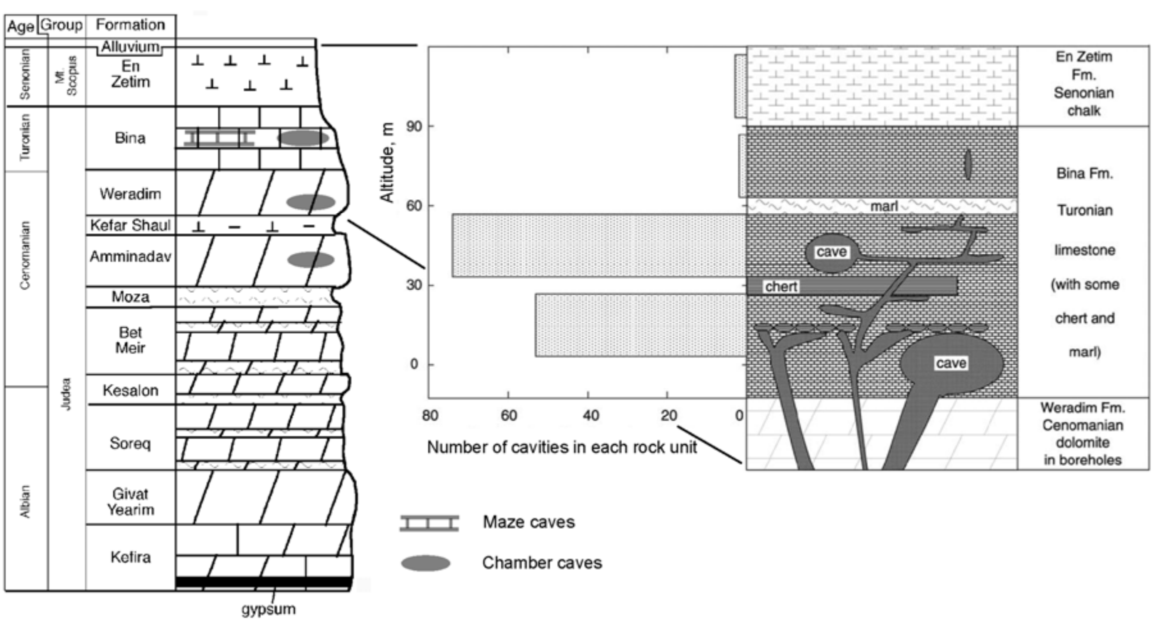

Figure 35. Stratigraphic section of Cretaceous formations in the Samaria Mountains, Israel (left) and a quarry in the confined zone (right) showing distribution of caves. The Judea Group consists of karstified dolomite and limestone with thin-bedded marl intercalations; Menuha Formation, composed of chalk and marl, is a confining unit. The histogram with the quarry section represents an arbitrary sample of voids encountered by the quarry from 1997 to 1999, the best available approximation of actual void frequency in the ASA zone in the confined flow area. Combined from Frumkin and Fischhendler (2005) and Frumkin and Gvirtzman (2006).

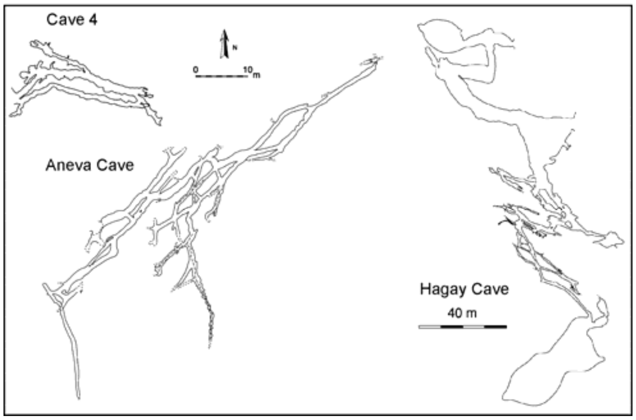

Figure 36. Maps of typical caves in the Turonian limestone. Network maze patterns commonly develop at two or three levels. Cave entrances are artificial. Broken lines on the Aneva Cave plan are rifts at the bottom of passages. Combined from Frumkin and Fischhendler (2005) and Frumkin and Gvirtzman (2006).

A few papers describing karst and caves in Saudi Arabia, Qatar and other regions of the Arabian Peninsula (*e.g.* Amin and Bankher, 1997; Peters *et al.*, 1990; Sadiq and Nasir, 2002; Hussain *et al.*, 2006) give strong evidence in favor of a hypogenic transverse origin of karst features, although not interpreting them in this way. The vast regional multi-story (eleven aquifers) artesian system comprises alternating sulfate, carbonate and clastic beds within the Mesozoic and Cenozoic Arab, Hith, Silaiy, Aruma, Umm Ar Radhuma, Rus, Dammam and Hofuf formations. This offers extremely suitable conditions for transverse speleogenesis. Numerous caves are mainly fissure- and slot-like passages or clusters of passages ("ghar" caves); some are clear rectilinear mazes. None of the caves show any genetic relationship with the surface. The regional artesian system discharges via numerous springs at the Gulf area, many of them being vertical pits ("ayns") through which groundwater rises from horizontal passage clusters at the base (Hötzl *et al.*, 1978). Amin and Bankher (1997) and Sadiq and Nasir (2002) implied that karst and caves in the area were formed epigenetically during past humid epochs of the Plio-Pleistocene. Hussain *et al.* (2006) suggested that network caves in the Upper Miocene calcareous sandstone in the Jabal Al Qarah area (Figure 37), which are being truncated now by denudational lowering, have developed due to weathering and enlargement of the fracture systems. The photographs in this paper show, in fact, very characteristic morphologies of confined maze caves, now in the process of un-roofing. It is argued here that, based on available information about regional litho- and hydrostratigraphy, hydrogeology, and cave patterns and morphology, the dominant mode of karst development in that region is probably hypogenic transverse speleogenesis. Numerous deep collapse sinkholes described in the region (*e.g.* Amin and Bankher, 1997; Sadiq and Nasir, 2002), some with unexplored caves at the base, are clearly related to regionally operating contemporary artesian speleogenesis rather than to presently inactive epigenic cave systems formed during past epochs of humid climates, as commonly assumed for the region.

A hypogenic transverse origin could be assigned to Magharet Qasir Hafeet Cave in the Jebel Hafeet ridge in the United Arab Emirates, described by Waltham and Fogg (1998). The cave has rift-like passages at depths of almost 100 m, connected to the surface through a series of vertical joints and shafts of apparently rising morphology. It occurs at the crest of an eroded anticline, in limestones that were confined by a clay-marl sequence in the past. Although initially a conventional phreatic origin was suggested for this cave (Waltham and Fogg, 1998), the possibility of *per ascensum* hydrothermal origin has been later acknowledged (Waltham and Jeannin, 1998).

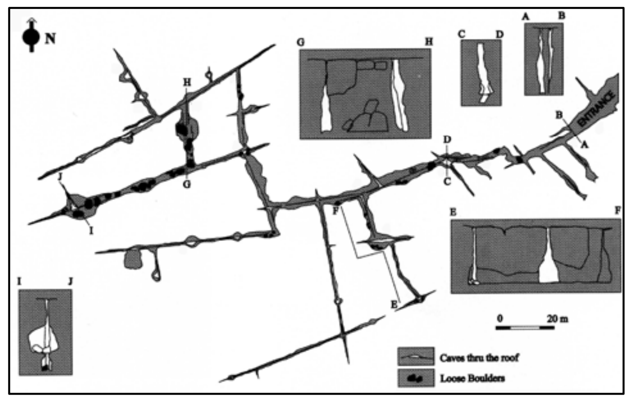

Figure 37. Plan view and cross-sections (insets) of Jabal Al Qarah caves formed in the calcareous sandstone of the Upper Miocene Hofuf Formation, northeastern Saudi Arabia (adapted from Hötzl *et al.*, 1978, and Hussain *et al.*, 2006).

Africa

Hydrothermal caves in massive Cretaceous limestones, including complex 3-D and network maze systems, are reported from northern Algeria (Collignon, 1983, 1990) and northern Namibia (Martini and Marais, 1996). Quinif and Dupuis (1989) described the artesian karst system of Ziaka in Zaire. Ziaka Cave is an emergence of the system explored by divers. It has the characteristic morphology of confined caves with numerous cupolas. The piezometric surface of the aquifer is higher than the base level of the Kwilu River.

Maze caves are common throughout the Transvaal Basin in South Africa, in the carbonates of the Malmani Subgroup of late Archaean age (2.5-2.6 billion years). Sterkfontein Cave is the best documented example (Martini *et al.*, 2003), well known for the hominin fauna found in a breccia fill. It is also an informative example of long but rather simple post-confined evolution of a maze cave. Sterkfontein Cave, together with the adjacent Lincoln and Fault caves, total 5.23 km in length, and form a complex, densely packed network of passages in cherty dolomite over a restricted area of 250x250 m (Figure 38). Twenty-five entrances open to the surface from near the top of a hill, the result of intersection of the cave by denudation lowering. The cave straddles the boundary between the basal Oaktree Formation and the overlying chert-rich Monte Christo Formation with distinct oolitic beds at the contact. The strata dip about 30° to the northwest and the cave layout shows the overall stratigraphic control.

In the chert-poor Oaktree Formation (the greater part of Sterkfontein), the passages are mainly of the fissure type. They reach heights of 15 m while the widths are on the order of a few meters. Large chambers can form by dissolution of partitions separating swarms of tightly spaced passages. Passages are often superimposed, adding more complexity to the map. Martini *et al.* (2003) report interesting observations of the original joints controlling passage development on the chert ceilings, which is unaffected by karst dissolution. The recorded width of these cracks varies from fractions of a millimeter to one centimeter. In the up-dip parts of the system where passages extend to the base of the chert-rich Monte-Christo Formation, they are broad and low, sandwiched between chert intercalations. In the down-dip direction, passages retain the same fissure-like morphology down to and below the water table over a vertical range of about 50 m. Caves extend to a greater depth elsewhere in the area, for instance to 79 m below the water-table, as observed by exploration after artificial de-watering of the aquifer (Moen and Martini, 1996).

Sterkfontein is a typical cave of the karst of the Transvaal basin, characterized by a deficiency of surface karst morphology but numerous network caves (Martini *et al.*, 2003). Another notable example is Wonderfontein Cave, a 9.4 km maze that is confined within an elevation range of only 3-4 m (Kent *et al.*, 1978). Martini *et al.* (2003) point out that the restricted extent of the easily penetrable passages, forming a dense network in plan view, is a characteristic shared by the majority of caves of the Transvaal Basin. This can be readily explained by the cluster nature of hypogene transverse speleogenesis.

Thirty years ago, when little was known about hypogenic speleogenesis, Martini and Kavalieris (1976) placed labyrinthic joint-controlled caves of the region in the hyperphreatic type, although specifics of this cave-forming mechanism were not really understood at that time. Apparently, "hyperphreatic" conditions are essentially confined conditions in cases of a stratified formation with vertical heterogeneities in permeability. In the recent work, Martini *et al.* (2003) guessed that these caves may have hypogenic origins, forming where deep water rose up and mixed with shallower flow systems close to the surface. This model was suggested by comparison with the caves of northern Namibia, where there is evidence of such upwelling for some of them (Martini and Marais, 1996). The confining cover in the Sterkfontein area was comprised of low-permeability shale and sandstone of the Ecca Formation (Permian), now retreated to the south. This suggestion, as well as characteristics of Sterkfontein described above, is fully compatible with the hypogenic transverse model.

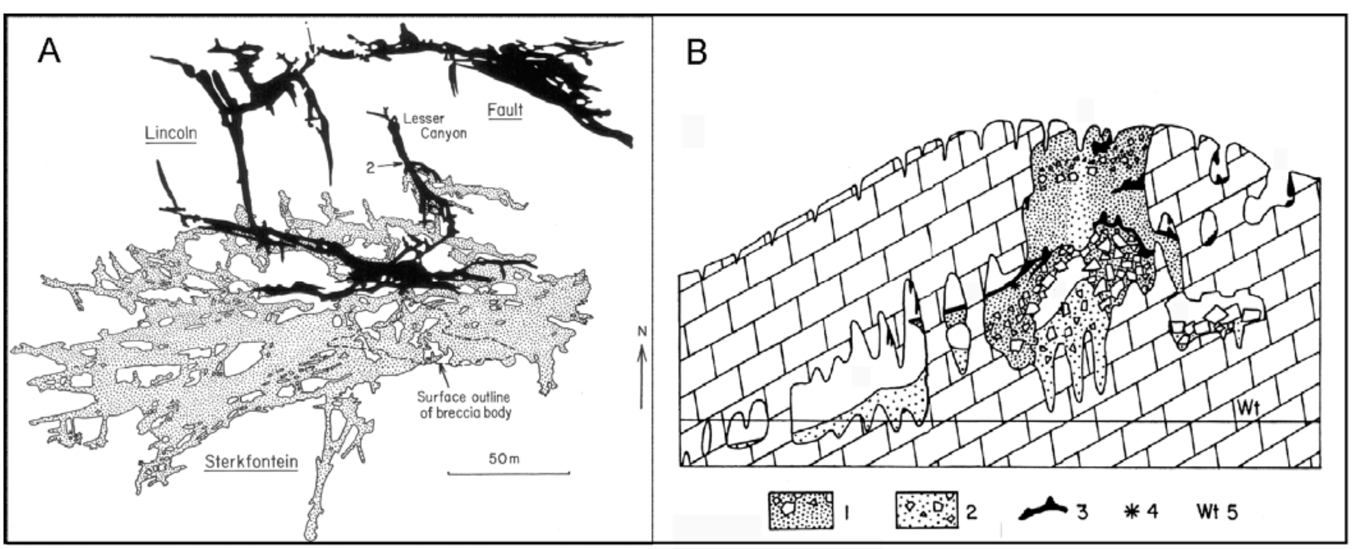

Figure 38. A = Simplified plan map of Sterkfontein Cave (gray, overlays and details omitted) and Lincoln-Fault Cave System (black). B = Cross-section of the Sterkfontein Cave showing unroofing of breccia body and breaching of the cave by denudation (adapted from Martini *et al.*, 2003).

The evolutionary scenario for Sterkfontein suggested by Martini *et al.* (2003) and slightly modified here in view of speleogenetic clarification, is as follows. The hypogenic transverse development of Sterkfontein could have started in the early Miocene, about 18 mya, following continental uplift and tilting of the African surface. The lack of cave fillings older than Upper Pliocene is explained by the presence of the protective confining cover during the African peneplanation. Considering the age of the oldest sections of the cave-hosted fossiliferous breccia, the breaching of the confinement might date to the end of the Miocene (~5 mya). The dating constraints for the *Australopithecus* bones found in the younger member of the breccia indicate that 3.3 mya the cave was already de-watered at 20-25 m above the present water table. The secular drop of the water table was irregular, comprising temporary rises, as evidenced by re-solution of calcified silt and breccia about 12 m above the present water table. It is remarkable that water table conditions lasting more than 3 My did not result in considerable modification of cave morphology and development of "water table" levels (see Figure 38-B).

North America

In North America, a hypogenic origin has been recognized for a number of caves, including such outstanding examples as the caves of the Black Hills and Guadalupe Mountains, but the true extent and role of hypogenic speleogenesis in this part of the world is still to be properly acknowledged. It is far beyond the scope of this work to provide a comprehensive and systematic review and re-interpretation of all cases where a hypogenic origin of caves was not previously recognized but can be suspected. Instead, only some instructive cases are mentioned, most of which are familiar to the present author through personal experience.

One of the earliest works that suggested a hypogenic transverse origin of caves in North America is an excellent study by Brod (1964; Figure 39) from eastern Missouri. Rectilinear fissure caves and small maze clusters are, developed along the bottom of the Ordovician Plattin Limestone by recharge from basal sandstones. These caves ascend to create a succession of pits and passages which breach the upper beds of varying lithologies to eventually provide focused discharge outlets for the artesian aquifer.

Outstanding examples of 3-dimensional (multi-story) network mazes are Wind and Jewel caves, some of the longest caves in the world, in the Mississippian Madison limestone in the Black Hills, South Dakota, USA (Figures 13 and 40). There are many smaller caves of this type in the area. Multiple stories in these mazes are stratiform, dipping in accordance with the stratal dip. The origin of the Black Hills caves is still debated (Palmer and Palmer, 2000b). Ford (1989) suggested the lifting maze model for the Black Hills caves, which is essentially a hypogenic transverse model. Bakalowicz *et al.* (1987) provided evidence for a hydrothermal origin of these caves and suggested that they were formed by thermal waters rising from the basal aquifer into the Madison limestones. Palmer and Palmer (2000b) suggest that mixing dissolution played a role in the main cave-forming stage and stress the importance of paleokarst zones in guiding cave development. The patterns and morphology of the Black Hills caves, however, display all the major features of confined transverse speleogenesis, the model being fully consistent with regional hydrogeologic settings and evolution.

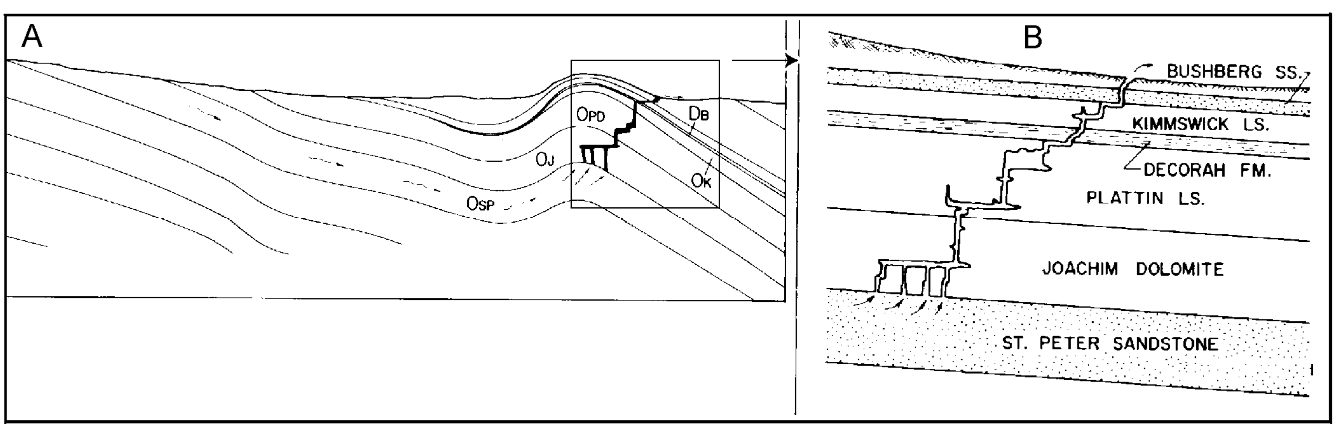

Figure 39. Fissure-like caves and ascending pits in eastern Missouri (from Brod, 1964).

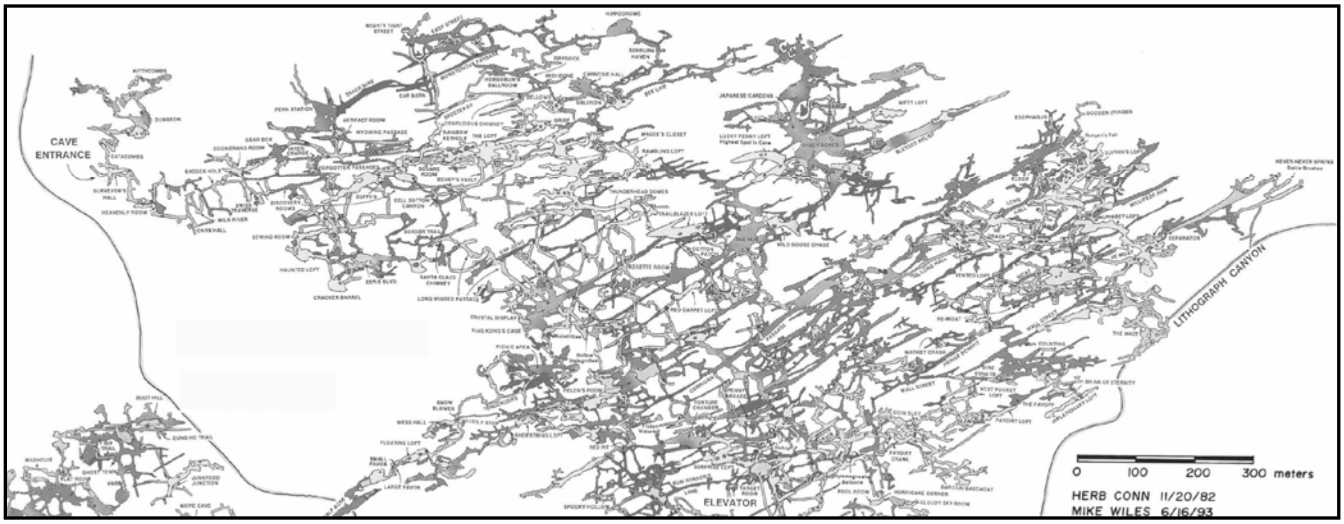

Figure 40. Map fragment of Jewel Cave, South Dakota, USA, showing superposition of different passage stories. Map by H.Conn and M.Wiles, courtesy of Jewel Cave National Monument. This is a type example of a multi-story maze pattern.

Mystery Cave in Minnesota is believed to be an example of a floodwater maze formed by a subterranean meander cutoff of a small river (Palmer, 2001; Figure 41). The cave is a 21-km long, relatively widely-spaced multi-story maze developed in sub-horizontal stratified Ordovician limestones. The cave has perfectly expressed the morphologic suite of rising flow as described in Section 4.4 (see Plates 1-E; 6-I; 7-C; 9-D and 14 for photographs of the cave's hypogenic morphology). Modifications due to epigenetic overprint are represented mainly by horizontal notching and are most concentrated in a few central passages (Plate 14, upper left photo). They do not override hypogenic morphology even in these major flow routes. The cave is a good setting for a detailed analysis of the overlap of epigenic development in a hypogenic cave.

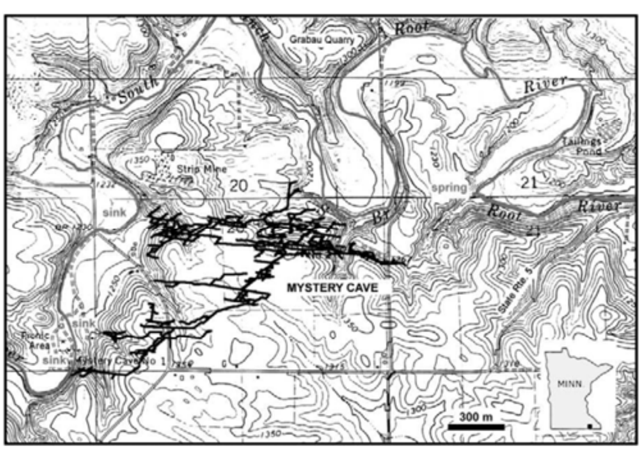

Figure 41. Mystery Cave, Minnesota, presumed to form as a subterranean meander cutoff of the West Fork of the Root River (from Palmer, 2001). However, the cave morphology strongly suggests a hypogenic origin (see Plate 13).

Several network maze caves and complex 3-D cave structures occur in Texas, developed in various sections of the thick stratified Cretaceous carbonates. Special examination of some of them, performed recently by G. Schindel and the author, strongly suggests their hypogenic transverse origin.

Robber Baron Cave is a network maze, occurring in several levels within a 16-m thick limestone interval of the Austin Chalk, a distinct unit in the upper part of the carbonate succession. The cave lies above the upper confining unit of the regional Edwards Aquifer. The mapped length of the cave is 1.33 km, probably only a part of the existing system (Figure 42; Elliott and Veni, 1994; Veni, 1989). One particular level is a master level, within which most of the passages are developed. The possibilities that are alternative to the confined *per ascensum* genesis - origin by diffuse recharge from above and by backflooding - can be definitely ruled out due to local conditions. The cave clearly displays all the features of the morphologic suite of rising flow (see Section 4.4 and Plates 2-C and 3-F). It was clearly formed under past confined conditions, probably due to rising flow from the presently confined Edwards. There are major springs in proximity to the cave, which discharge water from the Edwards through the Austin Chalk, a likely analogue of the past situation at Robber Baron Cave.

Figure 42. Robber Baron Cave, Texas, USA, a maze cave in Cretaceous limestone, above the confined zone of the Edwards Aquifer. Survey: San Antonio Grotto, 1976-77 (from Veni, 1989).

Amazing Maze Cave (the cave name) is a very densely packed 3-D network maze with 9 km of passages surveyed to date. It was originally opened by a road cut in a low-relief hill within a broad valley of Edwards Plateau outliers. Most of the passages lie at a single level but three or more other levels, connected by outlet/feeder structures, can be distinguished in the cave within a 22 m thick stratified section of the Fredericksburg Formation, Cretaceous Edwards Group (Figure 43; Elliott and Veni, 1994). The cave has no features indicative of downward percolation, considerable lateral flow or water table action. Instead, the morphologic suite of rising flow is perfectly represented in the cave (Plates 4; 9, F and H). Many passages on the master level contain massive bodies of microcrystalline gypsum, commonly coated with calcite crust. This gypsum is similar to the secondary gypsum known from the Guadalupe Mountains caves, indicative of a sulfuric acid dissolution mechanism. Gypsum masses in Amazing Maze Cave have numerous "vents" in them, which are morphologically continuous with feeders below. Some of these holes have gypsum blisters at the top. Gypsum masses do not fill passages from side to side but occupy central sections, leaving wide gaps between the gypsum body and walls. They were apparently formed under water-filled conditions. During our examination of the cave we found massive occurrences of endellite of intense purple color in several parts of the cave. In the main occurrence, endellite rims a feeder in a small passage at the upper level developed along a diffusely permeable bed (Plate 4-D). Identification of endellite has been confirmed by SEM and elemental analysis by M. Spilde (University of New Mexico). All of these characteristics strongly suggest the confined transverse origin of Amazing Maze Cave, in which, at least during some episodes, dissolution by sulfuric acid had contributed to speleogenesis through mixing of uprising H_2S-bearing waters with oxygenated waters flowing laterally through some more transmissive beds. The cave is located above the White and Backed Oil Field, and a source of sulfates could be evaporitic beds in the Trinity Group below.

The presence of the massive gypsum and endellite in a typical stratiform maze cave, which was apparently formed by rising flow and bears no signs of any water table development, is important for interpretation of these secondary formations in the context of sulfuric acid speleogenesis (see below for discussion of the Guadalupe Mountains). Both Amazing Maze Cave and Robber Baron Cave are representative, in morphological and geological respects, of a great number of maze caves throughout the United States, so that their identification as transverse hypogenic caves serves to establish this mode of speleogenesis as the dominant mechanism for the formation of maze caves (see Section 4.3 for discussion of the maze caves controversy).

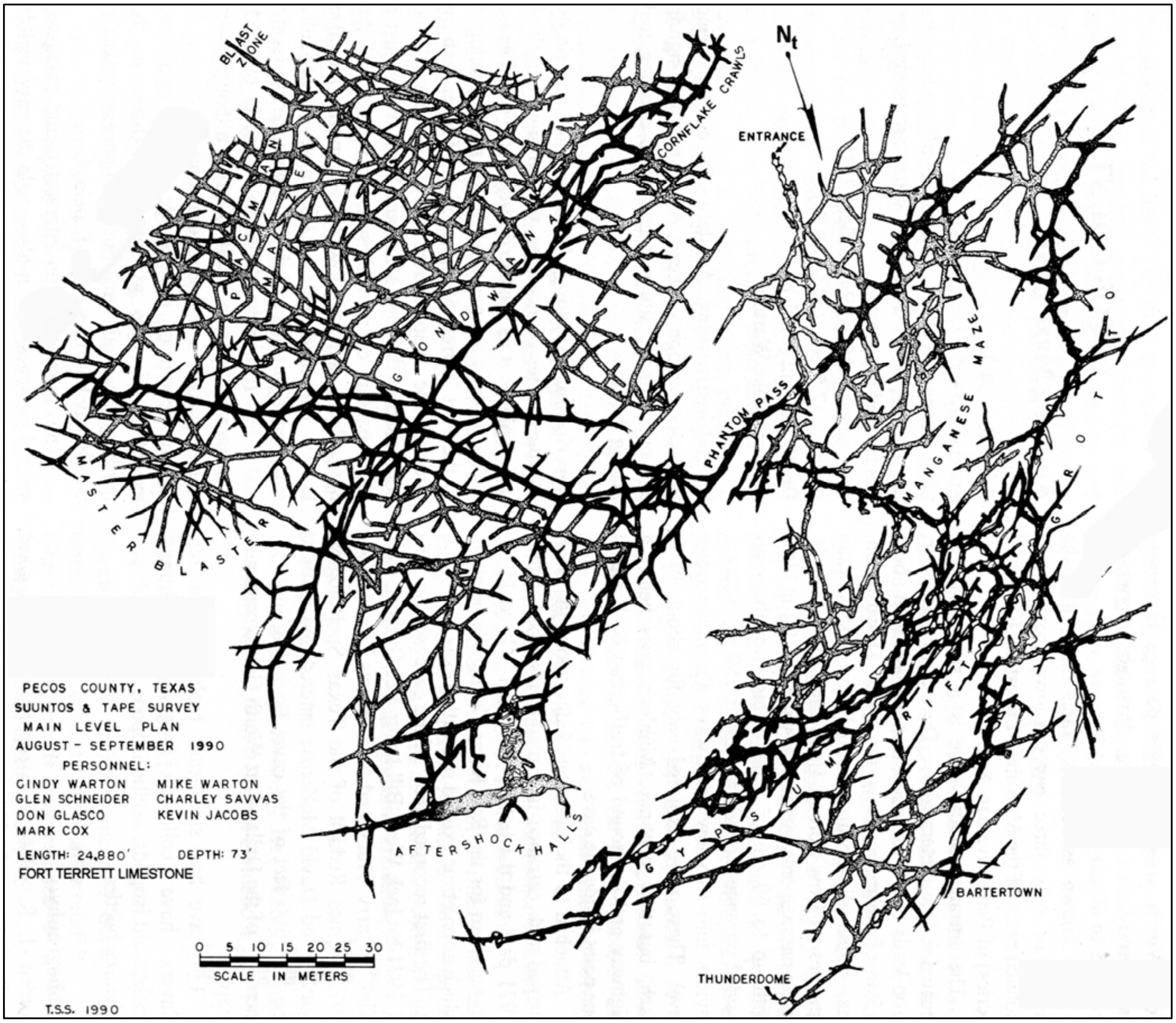

Figure 43. Amazing Maze Cave, Texas, USA, a multi-story maze cave in stratified Cretaceous limestone, an example of confined transverse speleogenesis in which dissolution by sulfuric acid took part (see text). Map courtesy of the Texas Speleological Survey (from Elliott and Veni, 1994).

The renowned Caverns of Sonora, located in the central portion of the Edwards Plateau within the drainage basin of the Devils River, is another instructive example of hypogenic transverse speleogenesis. The 2.3-km long cave consists of mazy stacks of nearly parallel joint-controlled passages stretching along two main trends (Figure 44), developed on four distinct stratigraphically-conformable stories within the vertical range of about 35 m in the Segovia (Fort Lancaster) Formation of the Cretaceous Edwards Group (Kastning, 1983; Onac *et al.*, 2001). The lower stories lie within massive dolomitic marly units and have generally larger dimensions than the upper stories. The upper stories with denser maze development lie within more porous beds separated by a marly unit. Cupolas at the uppermost story open up into a distinct bed of touching-vugs type porosity ("burrowed bed"), which probably served as a "receiving aquifer" during the ascending formation of the cave (Figure 9). Most of the cave lies beneath that vuggy bed, which is overlain by the thick unit of massive limestone that provides a caprock. Another maze cave in the vicinity, Felton Cave (2.05 km), is in many respects similar to Caverns of Sonora but has more diverse trends of passages (Kastning, 1983).

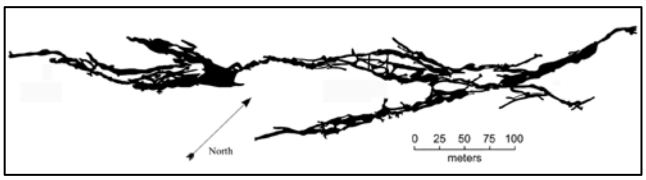

Figure 44. Map of Caverns of Sonora (by G. Veni and P. Sprouse; adapted from Elliott and Veni, 1994).

The cave perfectly displays the morphologic suite of rising flow (Plates 4-G, 6-B and 9 A, C and G). Passages at different stories are often co-planar, with rift-like or oval connections between stories. Smaller connections are domepits from the perspective of a lower story. Where passages at different stories are not co-planar, they are connected by steep passages with a rising sequence of dome-like forms. Ceilings, especially at upper levels, demonstrate complex half-tube/pendant morphology. The morphology of Caverns of Sonora bears strong imprints of dissolution by multiple buoyant currents and shows no appreciable modification by a water table or epigenic recharge.

Speleogenesis of Caverns of Sonora had been interpreted by Kastning (1983) in terms of a classic epigenic concept, assuming passage development by lateral flow recharged from above, with progressive shifting from upper levels to lower levels, in response to lowering of base levels. The ascending hypogenic origin of the Caverns of Sonora, besides morphological and hydrostratigraphic considerations, is strongly corroborated by the recent finding of metatyuyamunite, a uranium-vanadium mineral diagnostic of sulfuric acid dissolution (Onac *et al.*, 2001). The cave was formed under confined conditions in the mixing zone between deep-seated H_2S-bearing warm fluids and an oxygenated shallow flow system.

The origin of the major caves in the Guadalupe Mountains, New Mexico, USA, including some of the largest caves in the United States such as Carlsbad Cavern (43.2 km long, 315 m deep) and Lechuguilla Cave (193.4 km long, 490 m deep), is firmly attributed to sulfuric acid speleogenesis (*e.g.*, among others, Davis, 1980; Hill, 1987, 2000a, 2000b; Palmer and Palmer, 2000a; Palmer, 2006). The caves are formed in carbonate reef and backreef formations of Permian age, exhumed during several episodes of uplift (of which the Cenozoic is believed to be the main one) from beneath largely evaporitic sediments of the adjacent Delaware Basin (Figure 45). Most of these caves are developed near the reef-forereef contact in the largely massive Capitan Formation and the reef-backreef contact between the Capitan and prominently bedded Seven Rivers and Yates Formations (DuChene and Martinez, 2000), but some caves or parts of caves lie within the backreef succession. Caves are scattered along the mountain ridge, which plunges from southwest (from elevations up to 2767 m) to northeast (to elevations of about 1000 m) for about 70 km. Many of these caves have stratigraphically-conformable multi-story maze patterns, network or spongework, or both, but some caves display complex vertically extended 3-D structures that include maze and chamber elements at many loosely defined stories, and sub-vertical conduits connecting them (Figures 16, 17, and 46). The caves show no genetic relationships with the surface and fit most other criteria for ascending transverse caves (Section 4.1). It is apparent that Guadalupian caves, or their segments, utilized various kinds of original porosity available throughout different members of the rock succession, including syndepositional faults and fractures (Koša and Hunt, 2006), other syndepositional features such as teepees (Plate 16), paleokarstic cavities and zones, uplift-related discontinuities, and vuggy porosity. Depending on their nature and position within the geological structure, various porosity systems (and hence respective cave elements) can be distributed conformably within the stratigraphy or be discordant to the bedrock structure.

Although caves in the region have received much scientific attention during the last 30 years, speleogenesis in the Guadalupe Mountains still has many controversial aspects. A comprehensive overview of speleogenesis in the Guadalupe Mountains and discussion of relevant issues is clearly beyond the scope of this book. But this case is treated here more extensively compared to other entries in this section because the Guadalupes are a prime reference region of hypogenic speleogenesis, and interpretations of their speleogenesis are highly important in illustrating hypogenic processes.

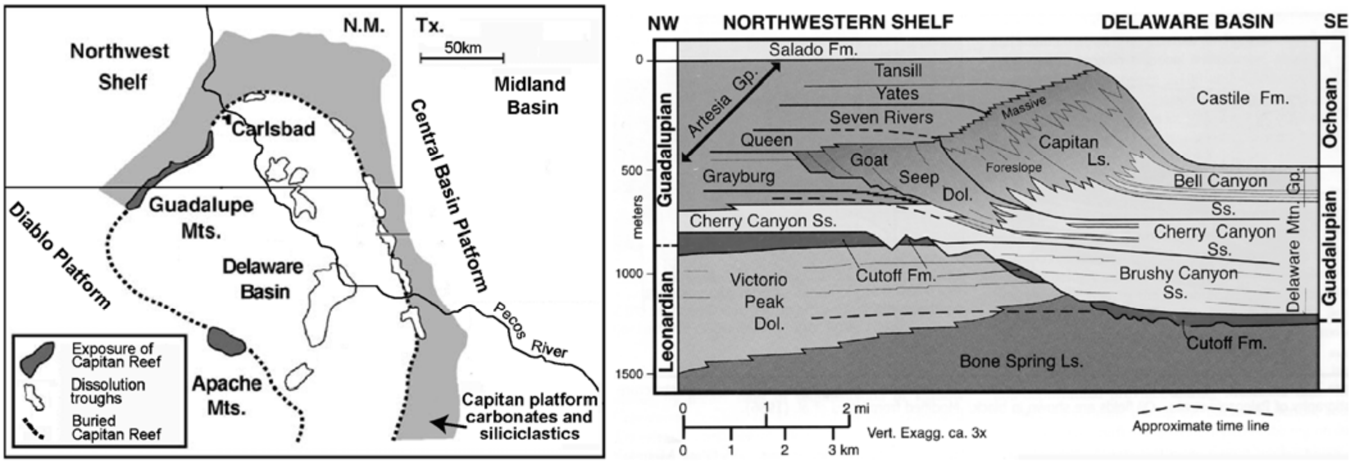

Figure 45. Regional structural setting of the Guadalupe Mountains (left; adapted from Koša and Hunt, 2006) and stratigraphic nomenclature of the Permian strata exposed in the Guadalupe Mountains (right; from Scholle *et al.*, 2004).

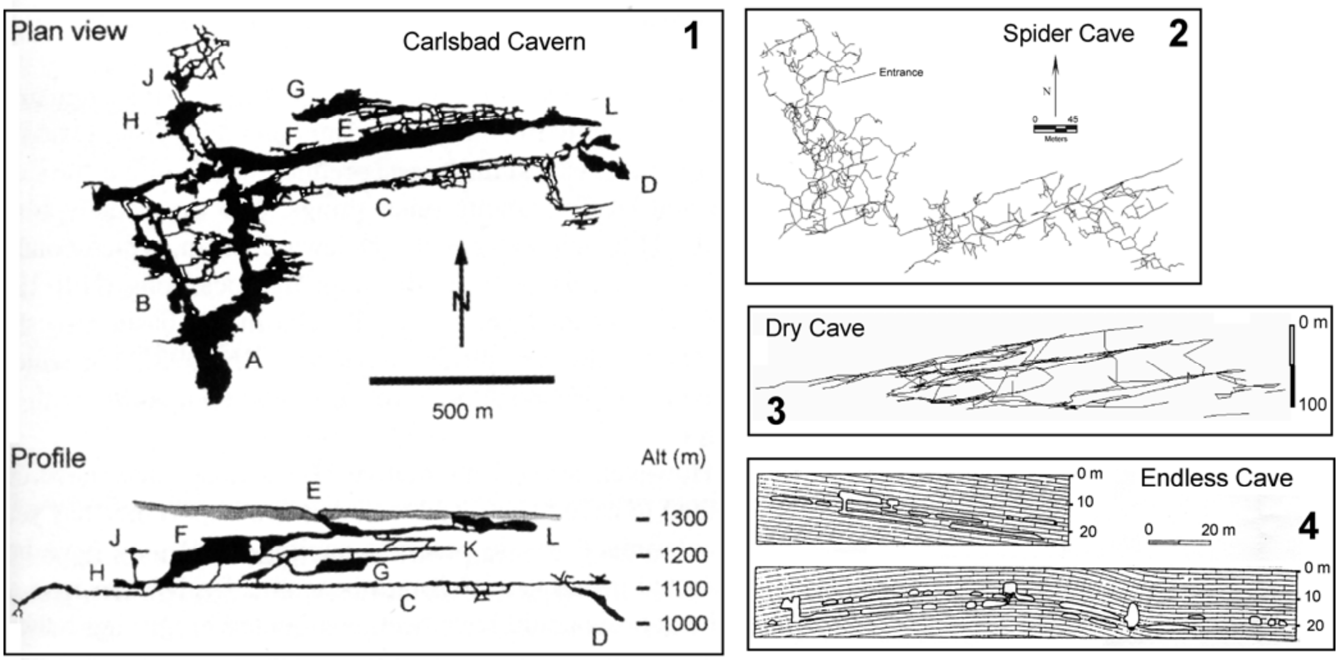

Figure 46. Plans and profiles of some Guadalupe caves: 1 = Carlsbad Cavern, plan view and profile (from Palmer and Palmer, 2000a); 2 = Spider Cave, plan view outline (by S. Allison, Carlsbad Caverns National Park); 3 = Dry Cave, profile outline (courtesy of Carlsbad Caverns National Park); 4 = Endless Cave, profiles and geology.

It is generally agreed, as a broad speleogenetic concept that water rich in H_2S rose from depth and reacted with oxygen at shallower levels within the reef/backreef formations to produce sulfuric acid (Jagnow *et al.*, 2000). Evidence for sulfuric acid dissolution is abundant, coming mainly from geochemical and mineralogical findings in various caves: massive gypsum deposits in many caves (Davis, 1980), isotopically-light sulfur in massive gypsum (Kirkland, 1982; Hill, 1987) and massive sulfur (Cunningham *et al.*, 1994), light-chain aliphatic hydrocarbons in sulfur and a number of sulfuric-acid related minerals, such as endellite, alunite, natroalunite, dickite, tyuyamunite, metatyuyamunite, aluminite and hydrobasaluminite (Hill, 1987; Palmer and Palmer, 1992; Polyak and Mosch, 1995; Polyak and Provencio, 1998).

Sulfuric acid as the main dissolutional agent in the Guadalupian speleogenesis seems to be almost universally accepted, although some researchers still cast doubt on whether it was the main cave-forming mechanism (*e.g.* Brown, 2006), and others point out that it may be difficult to separate the effects of sulfuric and carbonic acid dissolution in a mixing zone setting where CO_2 generated by carbonate dissolution is not allowed to escape from the cave-forming zone (Palmer and Palmer, 2000a). Although H_2S is firmly established to be the result of sulfate reduction processes involving hydrocarbons, its exact source is still debated. Hill's model (1987, 1996) implies that the gas migrated updip from the adjacent Delaware Basin from the Bell Canyon Formation, although this view is not well tied with paleohydrogeology. DuChene and Cunningham (2006) suggested the Artesia Group of the Northwest Shelf as an alternative source of H_2S. Another option is H_2S derived from deep source rocks below the Capitan reef (DuChene, 1986). The resolution of this issue will depend on a revised paleohydrogeological model, as most of the gas reached the cave-forming zones in aqueous, not in gaseous, form (Palmer and Palmer, 2000a). In addition, results of possible hydrothermal speleogenesis during the Miocene (Phase 3 caves of Hill, 2000b; 1996) would be apparently utilized and modified by later processes invoking sulfuric acid; the effects of these processes are again difficult to separate. Available radiometric dates for sulfuric acid footprints (alunite from various caves, 12.3-3.9 myr; Polyak *et al.*, 1998) certainly post-date the main phase of cave formation for respective caves and do not necessarily indicate that it was related to sulfuric acid.

However, the primary problem is that the paleohydrogeologic environment for speleogenesis of the Guadalupean caves is not well-discerned. Most works on speleogenesis in the Guadalupe Mountains have been focused on geochemistry, mineralogy, and speleothems but have largely left out in-depth studies of the cave-forming hydrogeologic environment. The notable exceptions include the paper by Palmer and Palmer (2000a), which provides, in addition to a sound hydrochemical background, a hydrogeological discussion based on regional morphogenetic analysis of cave patterns and morphology. DuChene and Cunningham (2006) discussed paleohydrogeological conditions based on analysis of tectonic/geomorphologic history.

The principal controversy is about whether the main cave development occurred under phreatic (bathyphreatic) conditions or was caused by dissolution at the water table and/or due to subaerial processes such as condensation corrosion, involving H_2S oxidation in water films and limestone/gypsum replacement. The only possibile resolution of this controversy lies in a systematic genetic analysis of cave patterns and meso-morphology, coupled with paleohydrogeological and paleogeomorphological analysis, and proper comparison with the broader context of hypogenic caves.

Most researchers view caves in the Guadalupe Mountains as a result of combined bathyphreatic and water table development. According to the original model (Davis, 1980; Hill, 1987, 2000a, 2000b), bathyphreatic (deep-water phreatic, rising flow) conditions were responsible for the strong vertical development of these caves, and for the formation of vertical tubes, fissures and pits; and water table (shallow-water phreatic) conditions were responsible for the horizontal development of caves along certain levels (corresponding to past regional base levels).

Palmer and Palmer (2000a) assigned most cave origins in the Guadalupe Mountains to bathyphreatic conditions, due to convergence of oxygenated water with deep-seated rising flow at depths up to 200 m below the water table or deeper. They acknowledged that the morphology of complex 3-D caves, such as Lechuguilla and Carlsbad Cavern, demonstrate rising flow patterns in both meso- and mega scales, from major feeders at the lowermost parts of the systems (such as the Rift and Sulfur Shores areas in Lechuguilla, and the Nicholson Pit and Lake of Clouds in Carlsbad Cavern) to highest outlet passages in the uppermost parts, including present entrances that served as outlets for rising groundwater (Figure 47). Different levels of the caves are connected by ascending passages, which show strong evidence for having been formed by rising aggressive water. Palmer and Palmer (2000a) further suggested that ascending complexes were formed in one stage, although they reserved the view that some chambers may post-date the systems, being enlarged at discrete episodes of water table development. An ascending flow pattern for the Guadalupe caves was also discerned, based on morphological observations, by Davis (1980). Observations by the author of this book in various caves of the region strongly support the views about their ascending transverse origin. In the meso-scale, the continuous succession of feeder-outlet and transitional features (MSRF, see Section 4.4) can be clearly traced throughout different levels and between them within the whole vertical range of caves. The sets of photographs illustrating MSRF components from various hypogenic caves (Plates 1-9, 11) include many examples from the Guadalupian caves. Overall, large and complex 3-D caves of the Guadalupe Mountains give compelling morphological demonstrations of an ever-ascending flow pattern.

Arguments toward the major role of water table dissolution and condensation corrosion (*e.g.* Hose and Macalady, 2006) are chiefly based on 1) comparison with active H_2S caves elsewhere in the world, 2) references to horizontal levels in caves, 3) references to gypsum and sulfuric acid-related minerals, and 4) references to specific morphologies. In addition, Palmer (2006) shows that the requirements for very low pH to form alunite found in the Guadalupe caves can be met only in subaerial conditions. In apparent contrast to his previously cited view, Palmer (2006) concluded that much, if not most, of the caves' volume, including the passages that ascend to entrances, has been produced subaerially due to sulfuric acid dissolution through absorption of H_2S by water films condensed on walls and ceilings. The above-mentioned arguments in favor of the major speleogenetic role of the water table and vadose development are briefly addressed below.

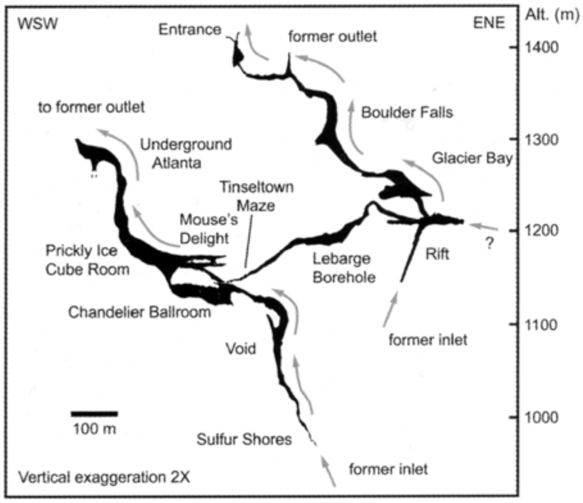

Figure 47. Projected vertical profiles through part of Lechuguilla Cave, showing the nearly independent flow systems through the entrance series and through the Sulfur Shores – Underground Atlanta systems (from Palmer and Palmer, 2000a).

1) The situation with reference to active H_2S caves that are currently at the water table stage is similar to the interpretation of maze caves of non-sulfuric acid origin (see Section 4.3); it is tempting to extrapolate present locally observed processes to more general interpretations of cave genesis. Commonly-cited examples include Kane Caves (Wyoming, USA), caves of Frasassi Gorge (central Italy) and Cueva de Villa Luz (Tabasco, Mexico). They do demonstrate quite aggressive subaerial dissolution by sulfuric acid (through H_2S oxidation by atmospheric oxygen), both in flowing water and in condensed water films, which certainly contributes to the morphology of these caves. This is insufficient, however, to generalize this as a major speleogenetic mechanism, responsible for the origin of a wide category of hypogenic caves. Mass balance considerations and the requirement for removal of dissolved matter (not met in many occasions, especially in extensive maze caves with diffuse outflow) make such generalizations unfeasible. Other caves in the same regions (or inactive parts of the same caves) include situations that are a poor fit to the water table/subaerial speleogenesis model (*e.g.* see the above description of hypogenic caves in central Italy).

2) References to horizontal levels in caves often include stories that are only somewhat horizontal (see Figure 23 and description above for Frasassi Gorge caves). Such stories do not correlate throughout adjacent caves and even between different areas of the same caves, which would be expected for true water table levels. They are commonly conterminous either vertically or laterally with clearly inclined stories within the same cave or in other caves of the same areas. In most cases, such stories are controlled by the distribution of initial porosity structures, either stratigraphically concordant (in most cases) or discordant to bedding. More discussion of "horizontality" is given below with regard to the Guadalupe Mountains caves.

3) Buck *et al.* (1994) described five types of gypsum in Guadalupe caves, two of which are basic and most relevant to the issue under discussion: (a) massive subaqueous gypsum sediment that forms large bodies in passages and rooms, and (b) subaerial gypsum crusts that replace bedrock by sulfuric acid reaction. Thin replacement gypsum crusts definitely form above the water table, but these are volumetrically insufficient to account for the caves' development. Massive gypsum sediments are found in many caves that show no signs or possibility of water table development, such as Monte Cucco caves in central Italy, Amazing Maze Cave in Texas, and Yellow Jacket, Spider, Dry, and Endless caves in the Guadalupe Mountains. This gypsum is apparently formed in water-filled passages. As to sulfuric-acid related minerals, these obviously reflect water table conditions due to the low pH requirement (Palmer, 2006), but this does not support a high relative significance of this environment in overall speleogenesis.

4) Some morphologies of the active H_2S caves are commonly referred to as being specific to sulfuric acid caves of water table/subaerial origin. Such references include maze patterns, abrupt changes in morphology, numerous dead-ends, ceiling cupolas, etc. However, these morphologies are specific neither to the water table situation nor to sulfuric acid speleogenesis. Instead, these features are characteristic of a wide class of ascending transverse caves, as argued throughout this book using both theoretical reasoning and references to caves, for which water table development and sulfuric acid dissolution are definitely ruled out.

The author's contention is that caves in the Guadalupe Mountains were mainly formed by rising transverse speleogenesis, possibly by both hydrothermal dissolution and sulfuric acid dissolution, with water table development playing secondary, modifying roles. Previous bathyphreatic hypotheses (Hill, 1987, 2000a, 2000b; Palmer and Palmer, 2000a) operated with the notion of an unconfined aquifer in the Capitan platform when discussing rising flow, *i.e.* an aquifer in which the water table under atmospheric pressure forms the upper boundary. However, speleogenesis in the Capitan platform prior to its exhumation and erosional lowering of the adjacent basin surface occurred under confined conditions. It is assumed by many workers that Salado evaporites and younger sediments spread across the shelf regions (Crysdale, 1987; Garber *et al.*, 1989; Ulmer-Scholle *et al.*, 1993; Scholle *et al.*, 2004; DuChene and Cunningham, 2006), so that they provided a confining cover over the carbonate platform before they were removed during

Tertiary uplift and erosion. It is also important to recognize that even after the cover's removal, a considerable confinement for rising flow through the Capitan complex, including the stratified backreef, was maintained due to the heterogeneous nature of the sequence in which both layered and discontinuous classes of heterogeneities are well-expressed.

Despite a general adherence to the unconfined aquifer notion, Hill (1996) refers to accumulation of H_2S in the Capitan reef in structural and stratigraphic traps, which implies substantial confinement of a gas-transporting flow. Palmer and Palmer (2000a) noted that stratigraphic trapping of rising H_2S water (confinement in terms of hydrogeology – A. K.) at and near the crest of an anticline accounts for the dense concentration of caves under McKittrick Hill. Davis (1980) inferred regional speleogenesis in terms of rising flow under pressure, and pointed to an analogous situation existing at the northeast end of the Guadalupes where the Capitan complex dips beneath the Pecos Valley. This implies confined settings.

Koša and Hunt (2006) provided a detailed study of syndepositional deformation in the Capitan Platform and demonstrated their role in speleogenesis. They showed that most faults and fractures are not cutting across the entire platform thickness, but terminate upward at some formational boundaries (Figure 48). This study illustrates well that various structures of initial porosity are confined to certain strata or otherwise distinct horizons (not necessarily stratigraphic) within the rock succession. Our observations in Yellow Jacket Cave and nearby outcrops further illustrate this important feature (Plate 16). This is the primary cause for cave stories in 3-D systems or entire caves to be restricted to certain horizons, either stratigraphically controlled or discordant to bedding. Poor vertical connectivity of initial porosity structures (fractures, faults and porous horizons), separated by less permeable massive beds of limestones or non-soluble rocks (such as low-fractured, dense siliciclastics; see Plate 16-3) creates multiple confining intervals for rising flow within this heterogeneous succession. The general importance of vertical heterogeneity across sedimentary sequences for confined hypogenic speleogenesis has been discussed in Section 3.4.

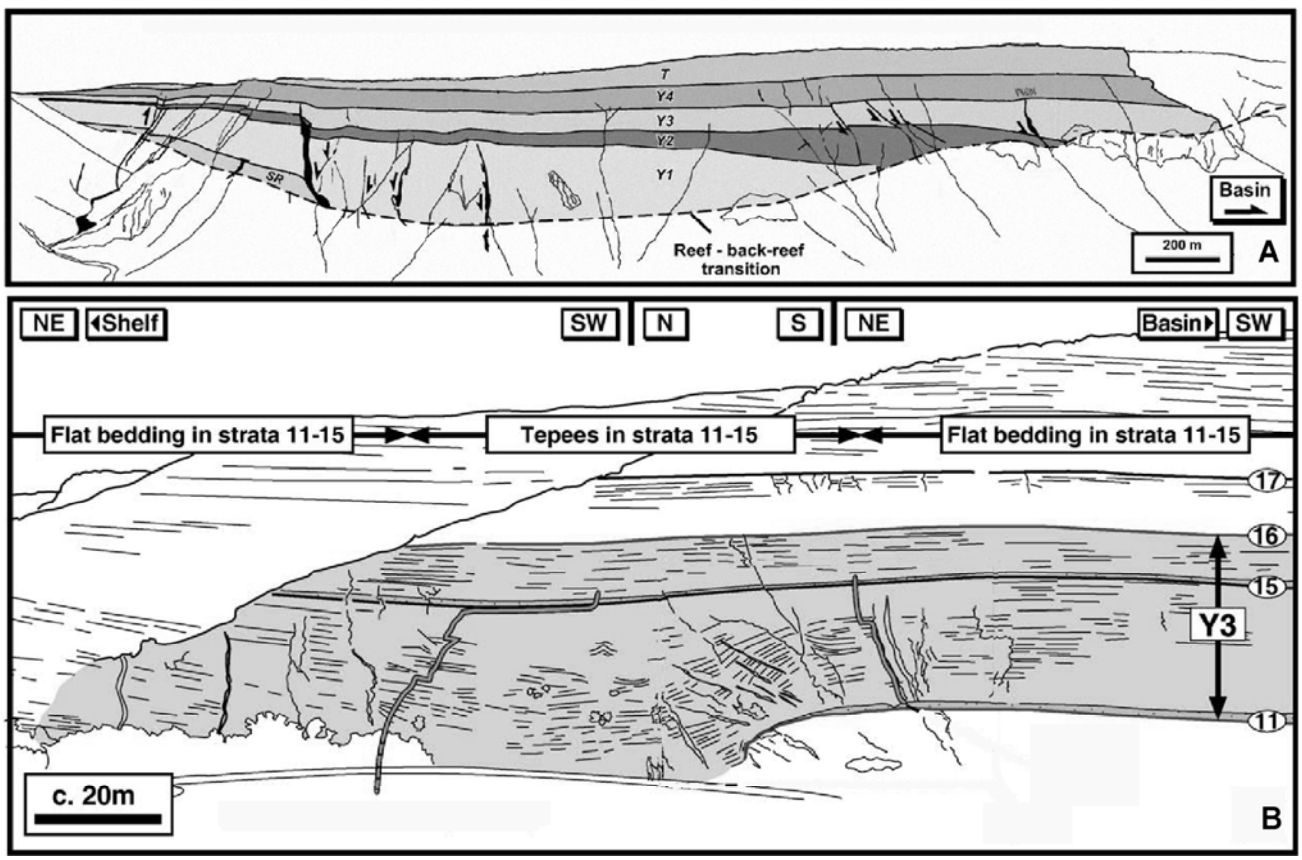

Figure 48. Line drawing of the geologic section through the Capitan Platform on the eastern face of Slaughter Canyon (A) and an exposure at Indian Shelter in Walnut Canyon (B), Guadalupe Mountains, NM. The drawing shows distribution of syndepositional faults and fractures, and other porosity elements. Note that most faults terminate upward at some formational borders, and that many sub-vertical ruptures occupy certain elevation horizons within the rock succession (from Koša and Hunt, 2006; see also other figures therein).

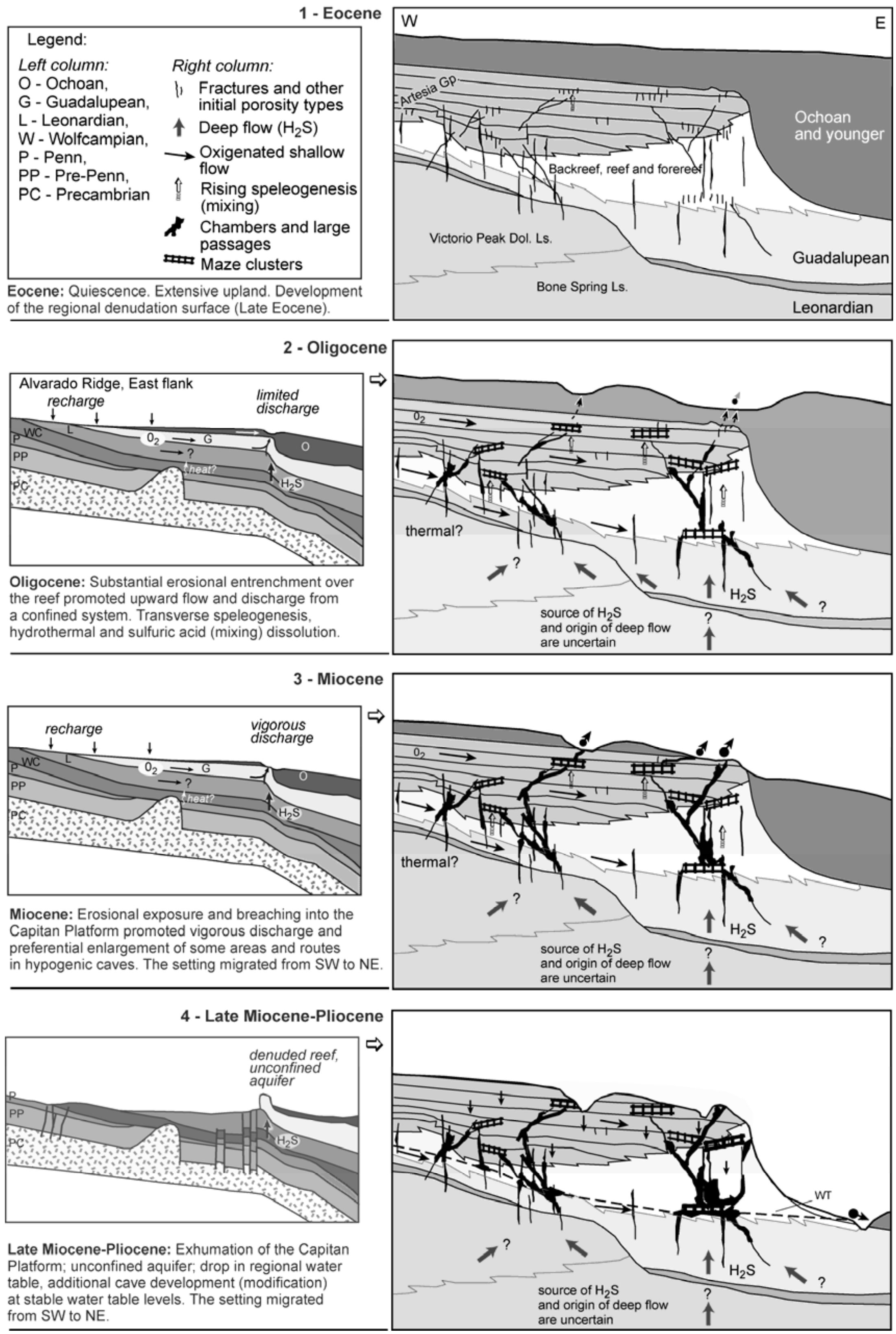

Figure 49. A proposed regional speleogenetic model for the Guadalupe Mountains. Diagrams of tectonic situation on the left are extracted from the broader tectonic profiles in DuChene and Cunningham (2006).

DuChene and Cunningham (2006) noted that previous work on the origins of Guadalupe caves were focused on the Guadalupe Mountains as a discrete block, rather than examining them in regional structural and tectonic context. They pointed out that the history of cave development in the Guadalupes is fundamentally tied to a regional paleohydrologic system, which developed in response to Laramide uplift of the Alvarado Ridge in New Mexico and west Texas. Their work provides an important background of the regional tectonic and geomorphic evolution, needed to decipher the paleohydrogeologic context of cave development. The speleogenetic model for Guadalupe caves suggested here is based on this evolutionary outline (Figure 49, diagrams in the left column, extracts from a broader picture of DuChene and Cunningham), and on the acknowledgement of vertical heterogeneity in initial permeability across the Capitan platform, a prerequisite for a confined aquifer system to develop (diagrams in the right column). The Alvarado Ridge began to rise in early Tertiary time, and by 38-35 myr, an elevated regional erosion surface extending across Colorado and New Mexico had developed. Prior to opening of the Rio Grande Rift, the ridge was an immense upland recharge area for aquifers that drained eastward into basins in eastern New Mexico and western Texas. The east flank of the Alvarado Ridge provided the potential for confined hydrodynamic flow through laterally-transmissive beds and horizons in the backreef and Capitan aquifers.

Initial erosional entrenchment over the platform, in response to either the uplift phase and/or climatic changes, established conditions for restricted discharge and hence for rising and convergence of shallower oxygenated flow from the westward recharge areas and the intermediate/regional deep flow systems (Figure 49, stage 2). As volcanism (Oligocene) and subsequent regional heating (early Miocene) imposed a substantial thermal gradient across the sedimentary sequence, speleogenesis could proceed through the hydrothermal mechanism ("Stage 3 thermal caves" of Hill, 1996, 2000a, 2000b). As discussed above, it is still largely an open question when H_2S began to enter the system, and from which source. The deep flow system could have originated from further upslope portions of the Alvarado Ridge, rising to the base of the Capitan platform from the Victorio Peak Formation or still deeper sediments, or from the Delaware Basin as suggested by Hill (1987). Palmer and Palmer (2000a) mentioned a possibility for a compaction/compression-driven flow system to rise periodically from the basin to deliver H_2S. It is plausible to assume that both hydrothermal and sulfuric acid dissolutional mechanisms operated, either simultaneously or sequentially, with alternating relative importance at different times through the main stage of rising transverse speleogenesis (Oligocene-Miocene). With the declining thermal gradient, sulfuric acid dissolution became the predominating mechanism, possibly overprinting much of the mineralogical evidence for the thermal contribution. The morphological suites of rising flow, extensively developed in the Guadalupe caves, contain strong imprints of a buoyant convection component in the morphology of already mature cave systems. As solute density variations are unlikely to be strong enough to drive buoyancy dissolution in the particular situation of the Guadalupe Mountains, the thermal density variations were probably the main cause for the free convection component, operative until the culminating phase of the confined development (see below). Vertical, inclined and quasi-horizontal elements of caves including multi-story maze clusters developed within a single although geologically quite prolonged, stage of rising transverse speleogenesis, being guided by distribution of respective initial porosity systems.

The culmination of this process occurred when erosion opened and locally truncated the Capitan platform, which caused vigorous discharge from the confined aquifer system (Figure 49, stage 3). The main distinction from the previous speleogenetic period was that rather pervasive cave development along all available paths changed to preferential development along select paths or zones connecting major feeders and ultimate outlets (rising springs). This phase was geologically short in each particular sector of the emerging Guadalupe Mountains, but probably added much of the volume to particularly large passages and rooms (*e.g.* Main Corridors in Cottonwood Cave and Carlsbad Cavern, the Big Room in Carlsbad Cavern, the Rift–entrance and the Sulfur Shores–Underground Atlanta series in Lechuguilla Cave; see Figure 47). It quickly changed to water table conditions (high initial position) in each emerging sector. The transition from a confined to unconfined situation began in the presently highest southwestern sector of the mountains (Guadalupe Peak), which was first to expose the reef from beneath the backreef confinement, and shifted in three main episodes northeastwards, as evidenced by the three distinct topography levels over the length of the ridge (DuChene and Martinez, 2000) and by progressively younger absolute dates from cave alunite (Polyak *et al.*, 1998; Polyak and Provencio, 2000; Figure 50). The alunite dates, ranging from 12 to about 4 My between the highest and lowest caves, record the water table episodes rather than the main speleogenetic development, so that they provide upper constraints on the transition (confined to unconfined) phases in respective sectors.

The water table situation (Figure 49, stage 4) migrated from the southwestern sector to the northeastern sector of the ridge through late Miocene-Pliocene. This speleogenetic stage resulted in the formation of sulfuric acid-related minerals and replacement gypsum, and increased the cave's volume due to water table and condensation dissolution. If stable water positions coincided with structurally-controlled stories, the latter

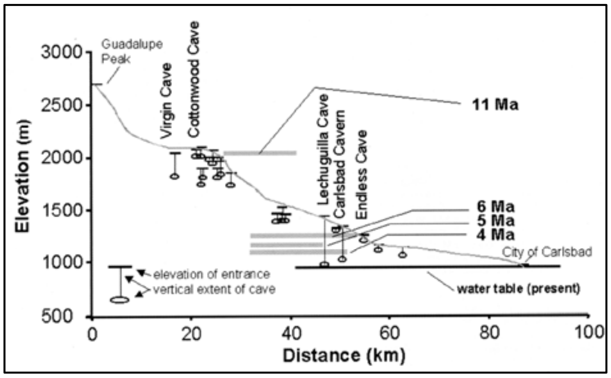

Figure 50. Longitudinal profile of the Guadalupe ridge (along the escarpment) from southwest (left) to northeast, with locations, elevations and vertical ranges of major caves. Age dates from alunite from four caves, and elevation of samples, are indicated (from Polyak and Provencio, 2000).

could receive better lateral integration. The development during the water table stage was unlikely to produce most of the cave volume as assumed by proponents of water table/condensation corrosion speleogenesis. There is no unambiguous evidence in cave patterns and morphology suggesting the prominent speleogenetic role of water table development. There are no truly horizontal levels in structural organization of most individual caves, nor noticeable correlation between quasi-levels in adjacent caves or in different parts of the same caves. There are no distinct, laterally continuous, marks in cave meso-morphology (such as horizontal notching, truncated partitions, etc.) even at those levels, which are assumed to be a result of water table development. In contrast, many parts of complex cave systems and individual caves demonstrate inclined stories of maze development (Figure 46), which apparently do not fit the water table concept. As shown by many examples throughout this book, such stories (including quasi-horizontal ones) are controlled by distribution of initial porosity structures. This is evident for stratigraphically concordant stories but is also an alternative to the water table) explanation for the bedding-discordant stories. The study of Koša and Hunt (2006) suggests that clusters of syndepositional fractures are often confined to certain elevation levels discordant to bedding (Figure 48; see also Plate 16). The quasi-levels in cave development are in many cases related to this control.

The total decline in the water table between the southwestern and northeastern sectors of the Guadalupe ridge is estimated to exceed 1000 m (Polyak *et al.*, 1998; DuChene and Cunningham, 2006), while the vertical ranges of individual 3-D cave systems vary from 20-30 to 250-490 m. Widespread correlation of cave stories within the water table concept would not be expected between caves scattered along the lengthwise direction of the ridge, as these caves experienced water table conditions at different times. However, it would be expected between caves for which elevation ranges overlap, located in proximity within the same transversal segments of the ridge. Palmer and Palmer (2000a) noted the lack of level correlations even between nearby caves and concluded that the confidence with which cave development (in the water table sense – A. K.) can be related to regional geomorphic events is limited. Instead, they suggested an elegant view in favor of the water table control on levels, namely that releases of H_2S from depth were episodic and occurred in different times and places. Horizontal levels were produced when these releases coincided with rather static water tables, so that bursts of cave enlargement occurred there. However, given that H_2S supply is associated with regional flow systems and events, it is unlikely that the gas releases occurred in such an individualized manner to caves located in close proximity, or only to particular major feeders within the same large caves. Other researchers argue that water table effects on cave development were not focused at particular levels but were distributed over the vertical range of caves due to water table fluctuations. This is certainly a sound possibility, but it gives no ground to claim the major speleogenetic role of water table development, as morphologic evidence of rising flow has not been overprinted by new, water-table related morphologies.

It follows from the above discussion that the origin of caves in the Guadalupe Mountains fits well with the broader class of ascending hypogenic transverse speleogenesis defined on a hydrogeological basis. The proposed refinement of the regional speleogenetic model is sufficient to explain virtually any features of cave patterns, morphology, and mineralogy observed in the region. The main speleogenetic stage of confined development probably involved both hydrothermal and sulfuric acid dissolution mechanisms and was quite prolonged in the geologic time scale. This discussion, taken within the overall context of this book, also suggests that caves of the Guadalupe Mountains, although being outstanding examples, are not unique, and that most of their characteristics (except geochemical and mineralogical) are not exclusive to sulfuric acid dissolution, as many works have suggested.

Central and South America

It is apparent from publications and exploration reports that Central and South America have a remarkable diversity of hypogenic karst. Detailed studies, however, are still scarce. Below only a few examples are referred to in order to highlight the diversity and some related speleogenetic issues.

Cueva de Villa Luz in Tabasco, Mexico, has received much attention during recent years as an example of active and dynamic, presently unconfined hypogenic development (Hose and Pisarowicz, 1999; Hose and Macalady, 2006). The cave is a 2-km long maze of passages lying almost horizontally 10-20 m below the surface, connected to it by numerous skylight entrances (Figure 51). The cave has more than 20 underground springs rising from conduits (up to 1 m x 2 m in width) in the floor and coalescing into a surface resurgence spring. Spring discharge (~270-300 L/s) is remarkably steady. Together with elevated temperature (28^{o} C) and high H_2S and CO_2 content of the water, this suggests a deep flow system possibly related to the magmatic system of El Chichon volcano, 50 km to the west. Springs inside the cave release H_2S and CO_2 into the cave atmosphere at concentrations that vary, respectively, from 1 to 120 ppm and from 0.03 to >3.2%. Replacement gypsum coats most walls throughout the cave. Fallen gypsum forms piles up to 1 m thick and is partially removed by the stream. The cave has many cupolas which are presumed to be actively growing due to chemical stoping. The origin of the cave is presumed to be largely due to subaerial dissolution at the stage of underground hypogenic springs (Hose and Macalady, 2006).

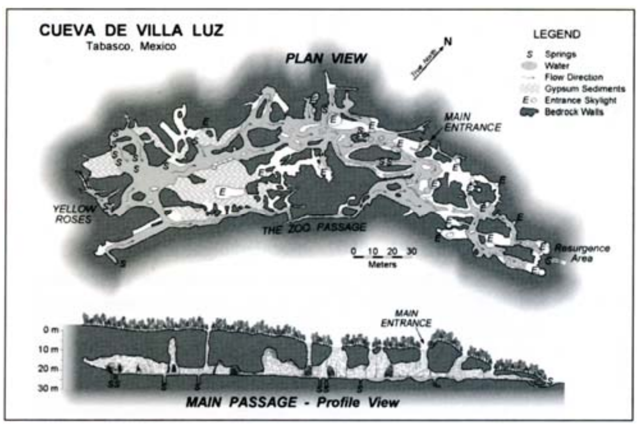

Figure 51. Map of Cueva de Villa Luz, Tabasco, Mexico (from Hose and Pisarowicz, 1999).

Sistema Zacatón (also known as Los Cenotes de Tamaulipas or Los Cenotes de Aldama) in the Sierra de Tamaulipas, northeastern Mexico, is one of the most outstanding examples of deep hypogenic karst, expressed as a series of extremely deep, phreatic megasinkholes, some of them being sealed with travertine precipitating at the water table (Figures 52 and 53, Plate 19 upper photo). Recently published results of the ongoing study of Sistema Zacatón provide a detailed description of the system and discussion of its origin (Gary *et al.*, 2003; Gary and Sharp, 2006). Besides megasinkholes – deep phreatic shafts – the system also includes ramiform, currently vadose cave passages, sub-horizontal phreatic conduits, broad overland travertine flows and relict spring flow travertine. El Zacatón is the deepest of the sinkholes, and the second deepest underwater shaft in the world with its measured depth of at least 329 m (dove to -284 m).

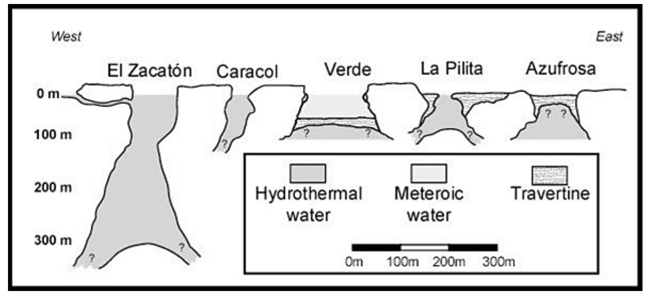

Figure 52. Profile view of southernmost sinkholes of Sistema Zacatón showing zones of varying water types and travertine morphology. The Verde pit is hypothesized to have a travertine floor formed during periods of low water-table levels (~50 m). This cap now serves as a flow barrier (from Gary and Sharp, 2006).

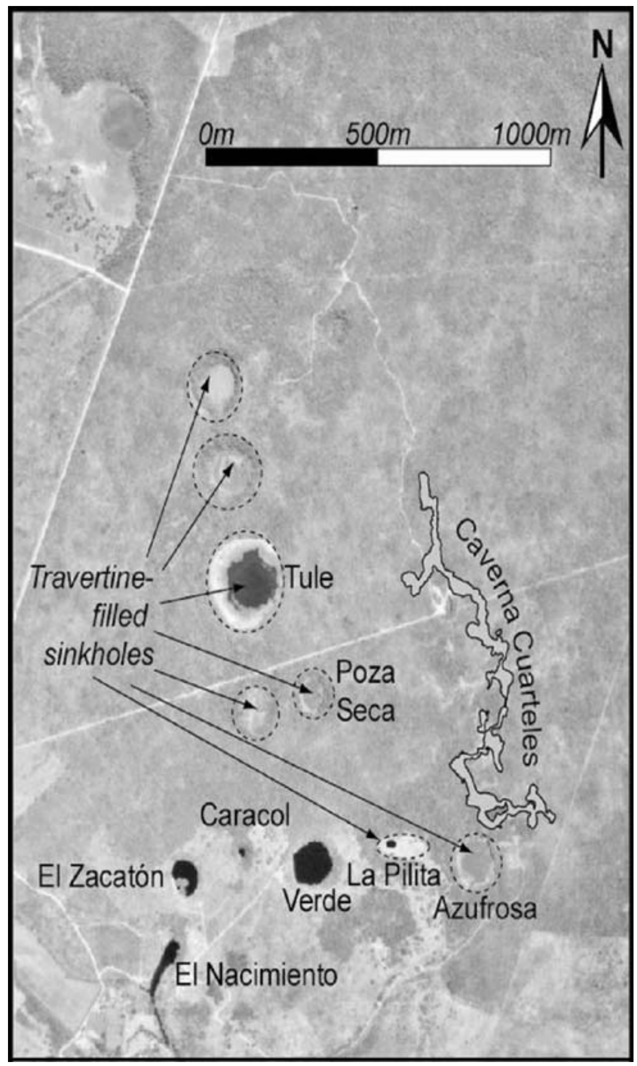

Figure 53. Aerial photograph of Sistema Zacatón showing the major features of the area. Numerous travertine-filled sinkholes are labeled and water-filled areas appear as black. Water flows underground from El Zacatón to the resurgence of El Nacimiento through a phreatic cave passage and then flows overland until becoming dispersed through broad honeycomb travertine deposits south of the area (from Gary and Sharp, 2006).

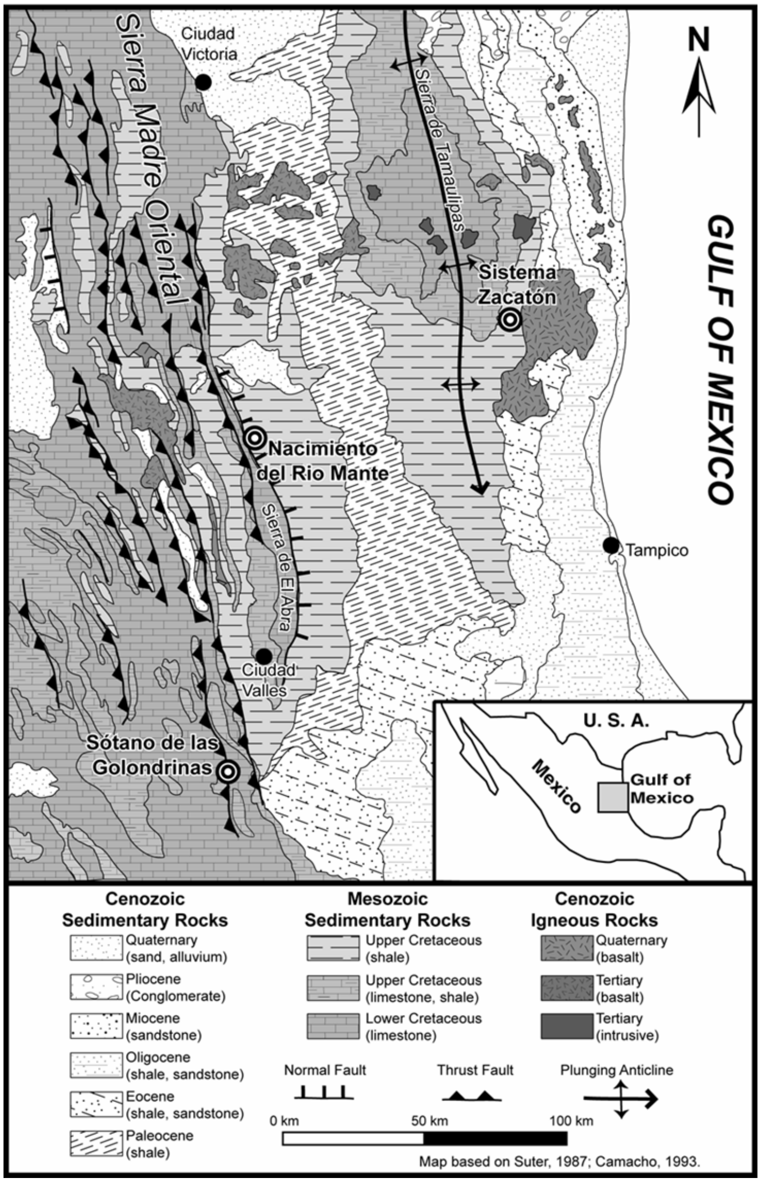

Figure 54. Geologic map of northeastern Mexico showing the location of the deep karst shafts in the region (shown as bull's-eye circles). Major depositional and structural features are also represented (from Gary and Sharp, 2006).

The system occurs in Lower Cretaceous forereef/reef/backreef limestones overlain by Upper Cretaceous argillaceous limestone and shale deposits (Figure 54) that cover much of northeastern Mexico. The Laramide orogeny and uplift exhumed the younger strata. During the Tertiary, igneous activity had a significant imprint on the regional geomorphology. The Villa Aldama volcanic complex, located within 5 km of El Zacaton, consists of Pliocene and Pleistocene lava flows and shield volcanoes, with the most recent igneous rocks dated at 250 ky (Gary and Sharp, 2006).

Sistema Zacatón is believed to have developed under the direct influence of Pleistocene volcanic activity, which provided the thermal gradient, CO_2 and H_2S to drive dissolution mechanisms at increased rates (Gary and Sharp, 2006). The inferred evolutionary model implies that the development of the karst system began much earlier, since a deep groundwater flow system with recharge at upland areas was established by the early Tertiary. It continued through the middle to late Tertiary under the influence of intermittent intrusive volcanic activity. Pleistocene volcanism accelerated and focused speleogenesis in the area.

Variations in connections with the deep flow system and in the degree of interaction with shallow groundwaters and surface waters account for varying geochemical characteristics of water in different sinkholes. Water in El Zacatón is undersaturated with calcite.

The primary trend of sinkholes/pits is roughly linear, north to south, coinciding with fractures observed in the area and the axial trace of the Tamaulipas Arch anticline (Figure 54). There is a secondary E-W trend in fracture and sinkhole pattern. El Zacatón's lateral extent and pattern of cavities at depth is unknown. It is also uncertain whether the shafts were formed due to collapses over large chambers at depth or as dissolution features of rising flow. Gary and Sharp (2006) believe that the sinkholes formed due to collapse. Another known deep phreatic shaft, 392-m deep Pozzo del Merro near Rome, Italy, shows the morphology of a rising shaft (Figure 27).

Numerous maze caves are known from several large basins in Brazil, formed in the predominantly carbonate Precambrian Una Group in the São Francisco Craton. Outstanding examples described from the Campo Formoso area are the 125 km long Toca da Boa Vista and 28 km long Toca da Barriguda caves, both developed in the carbonate sequence of the Salitre Formation (Auler and Smart, 2003). The caves show no genetic relationships to the surface, display many features characteristic of hypogenic caves and no vadose features. The cave plans exhibit densely packed, joint-controlled patterns, predominantly network, with some larger passages and chambers (Figure 55). The cited work mentions continuous phreatic dissolutional features that can be traced up to the cave entrances, and suggests that the cave passages once extended above their present surface elevation, being intersected by denudational lowering of the surface.

Auler and Smart (2003) suggested a hypogenic origin for these caves, but their connotation of "hypogenic" is somewhat controversial, based on the "source of acidity" definition of the term rather than on hydrogeological criteria. Referring to the lack of deep ascending passages that mark the vertical path of the acid source and the absence of volcanic activity and hydrocarbon deposits in the area to supply acidity from depth, they proposed that the caves were "*formed when oxic meteoric waters penetrating from former recharge zones percolated downwards towards the laterally flowing aquifer, coming in contact with pyrite contained within the carbonates*" (p. 165). For this dissolution mechanism to be a feasible option, it is necessary to show that pyrite could occur in considerably higher concentrations than commonly observed in the carbonate rocks. More importantly, the suggested cave-forming flow system is essentially epigenic and unconfined, implying recharge from the surface and lateral flow through the carbonates. Such a flow system is unlikely to produce extensive maze caves that have hypogenic characteristics and display no morphogenetic relationships to the surface. A possible alternative option is the hypogenic origin in the hydrogeological sense, with recharge from a basal unit. Ascending feeders to the master passages do not necessarily have to be prominent conduits but could be smaller features scattered through the network, as documented in many maze caves in other regions. It is noted in the cited paper that the original bedrock floor in these caves is almost always obscured by breakdown or sedimentation, so that structures of recharge from below may be present but not noticed. More detailed morphological studies and a consideration of litho- and hydrostratigraphy and the geomorphic evolution of the basin are needed to decipher the origin of these remarkable caves.

Figure 55. Toca da Boa Vista and Toca da Barriguda caves in Brazil (from Auler and Smart, 2003).

Australia

Osborne (2001b) summarized that many caves in Paleozoic limestones of eastern Australia have morphological, hydrological, and mineralogical features that suggest a hydrothermal or artesian origin. He pointed to many features that conform to the above criteria of ascending transverse speleogenesis, disregarding the possibility of origin due to downward recharge through the caprock, and concluded that they have developed by upward recharge from basal aquifers rather than by sinking meteoric waters. These caves are structurally guided, show maze patterns (Figure 56), have "halls-and-narrows" morphology (Osborne, 2001a), numerous dead-end terminations, blades, partitions (Osborne, 2003), roof pendants, cupolas and pockets. Previously such caves were described by the term "nothephreatic" (Jennings, 1985), with the meaning that they were excavated by diffuse flow under phreatic conditions, which is a close approximation to confined settings.

A remarkable example is Exit Cave in Tasmania consisting of over 40 km of network passages through which a major stream is captured underground. The captured stream flows through only some of the passages and was not responsible for the cave's development (Osborne, 2001b).

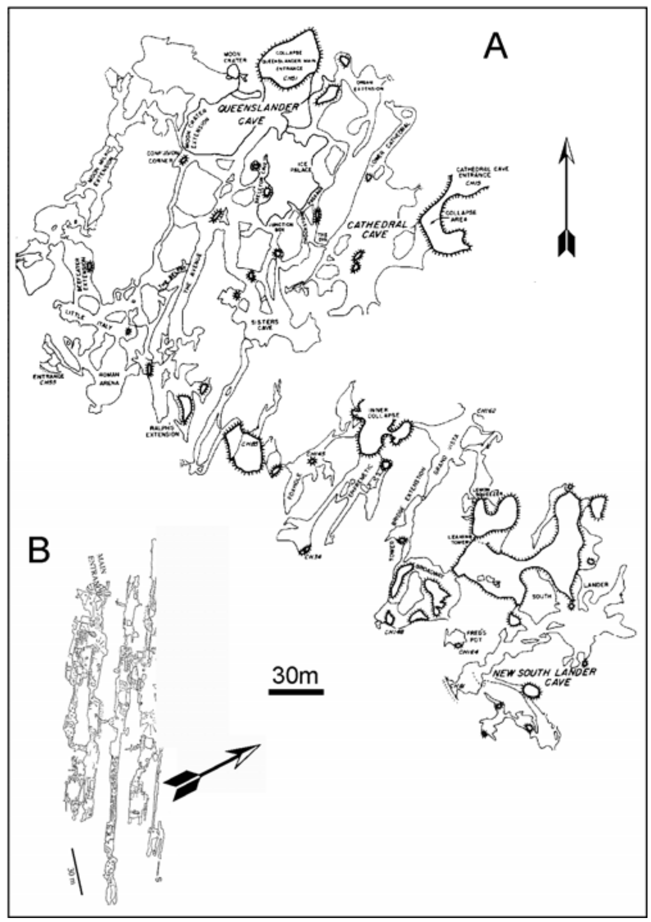

Figure 56. Typical hypogenic caves of Eastern Australia (from Osborne, 2001a). A = The Queenslander-Cathedral Cave System, Chillagoe, Queensland; B = Ashford Cave, New South Wales.

Synopsis

Several general points can be derived from the above regional overview.

1) Hypogenic speleogenesis is much more widespread that previously assumed. It is identified in various lithologies and geological and tectonic settings. Despite these variations, the resultant caves demonstrate a remarkable similarity in their patterns and meso-morphology, which suggests that the hydrogeologic settings were broadly identical in their formation. The morphologic suites of rising flow with buoyancy components are clearly identifiable in most of these caves.

2) In many areas more than one dissolutional process, such as CO_2-driven hydrothermal dissolution and H_2S (sulfuric acid) dissolution, is recognized to form hypogenic caves. They operated either simultaneously or sequentially, and it is often difficult to discriminate between their respective speleogenetic effects.

3) The great majority of accessible hypogenic caves are relict. Many of them bear signs of overprint by epigenetic processes. However, many active (in the sense of continued hypogenic development) caves are documented, either directly or indirectly. Active hypogenic caves are found in various current hydrogeologic environments: at the water table, at depth in presently unconfined aquifers (rising flow through them indicates vertical head gradient from still deeper aquifers and the hypogenic component), and in currently confined conditions. In many cases these settings (with regard to cave-hosting formations) occur proximal to each other in the same region, suggesting a speleogenetic evolutionary sequence in response to uplift and erosional lowering/entrenchment.

4) Many maturely developed hypogenic caves with type morphological characteristics are unambiguously shown to develop under confined conditions. None of the morphologically mature caves presently active at a water table were unambiguously shown to form in the respective contemporaneous settings. This suggests that confined settings are the principal hydrogeologic environment for hypogenic speleogenesis, which is in agreement with the broad analysis of hydrogeological evolution and the ascending transverse speleogenetic model. However, hypogenic caves may experience substantial modification under subsequent unconfined stages, especially when H_2S dissolution mechanisms are involved.

5) Whether or not water table/subaerial dissolution can be a major mechanism in also creating features that occur by hypogenenic processes remains an open debate and requires more research.

4.6 Comparison of confined versus unconfined conduit porosity

The distinctions between hypogenic (confined) and epigenic (unconfined) speleogenesis can be illustrated by the analysis of morphometric parameters of typical cave patterns. Klimchouk (2003b) compared two representative samples of typical cave systems formed in these two settings. The sample that represents unconfined speleogenesis consists solely of limestone caves, characteristically displaying branchwork patterns. Gypsum caves of this type tend to be less dendritic. The sample that represents hypogenic-confined speleogenesis consists of both limestone and gypsum caves that have network maze patterns.

Passage network density (the ratio of the cave length to the area of the cave field, km/km^2) is one order of magnitude greater in confined settings than in unconfined (average 167.3 km/km^2 versus 16.6 km/km^2). Similarly, an order of magnitude difference is observed in cave porosity (the fraction of the volume of a cave block occupied by mapped cavities; 5.0% versus 0.4%). This illustrates that storage in maturely karstified confined aquifers is generally much greater than in unconfined aquifers. Average areal coverage (a fraction of the area of the cave field occupied by passages in a plan view) is about 5 times greater in confined settings than in unconfined (29.7% versus 6.4%). This means that conduit permeability in confined aquifers is appreciably easier to target with drilling than the widely spaced conduits in unconfined aquifers.

The results in Tables 1 and 2 clearly demonstrate that there are considerable differences between confined and unconfined settings in the average characteristics of cave patterns and porosity. The fundamental cause of this difference in conduit porosity is demonstrated to be a specific hydrogeologic mechanism inherent in confined transverse speleogenesis (restricted input/output), which suppresses positive flow-dissolution feedback and speleogenetic competition in fissure networks (Klimchouk, 2000a, 2003a). This mechanism accounts for the development of more pervasive channeling and maze patterns in confined settings where appropriate structural prerequisites exist. In contrast, the positive flow-dissolution feedback and competition between alternative flowpaths dominates in unconfined settings to form widely spaced dendritic cave patterns.

Table 2 shows no appreciable difference of parameters between gypsum and limestone caves formed in confined settings. However, there are noticeable differences between parameters of particular caves even from the same region (Table 1). For example, compare the characteristics of Jewel and Wind caves, both occurring within the slopes of the structural dome of the Black Hills, or characteristics of the gypsum mazes in the western Ukraine.

There are two explanations for such differences. First, one of the implications of the hypogenic transverse speleogenetic model is that virtually all hydrogeologically

TABLE 1
Characterization of cave patterns and porosity in unconfined versus confined aquifers

Cave	Length, km	Area of cave, m^2x10^6	Volume of cave, m^3x10^6	Area of cave field km^2	Volume of rock, m^3x10^6	Specific volume, m^3/m	Passage density, km/km^2	Cave porosity, %	Areal coverage, %
"Common" caves – speleogenesis in unconfined settings									
Blue Spring Cave, Indiana, USA, Carboniferous limestones	32.0	0.146	0.5	2.65	119.34	15.6	12.07	0.42 (0.08)	5.5 (1.1)
Mammoth Cave, Kentucky, USA, Carboniferous limestones	550.0	1.386	8.0	36.78	3310.2	14.5	14.95	0.24 (0.09)	3.77 (1.4)
Friars Hole System, WV, USA, Carboniferous limestones	70.0	0.3	2.7	4.37	349.92	38.6	16.00	0.77 (0.28)	6.86 (2.5)
Krasnaya Cave, Crimea, Ukraine, Jurassic limestones	17.3	0.064	0.27	0.74	37.0	15.5	23.23	0.15	8.55
Maze caves – speleogenesis in confined settings									
Jewel Cave, South Dakota, USA, Carboniferous limestones	148.01	0.67	1.49	3.01	135.63	10.00.0	49.11	1.10	22.20
Wind Cave, South Dakota, USA, Carboniferous limestones	143.2	0.43	1.13	1.36	61.0	7.9	105.68	1.86	31.73
Knock Fell Caverns, N. Pennines, UK, Carboniferous limestones	4.0	0.006	0.012	0.02	0.12	3.0	170.94	10.26	25.64
Fuchslabyrinth Cave, Germany, Triassic limestones (Muschelkalk)	6.4	0.0058	0.007	0.03	0.15	1.1	217.61	4.80	19.55
Moestroff Cave, Luxembourg, Triassic limestones (Muschelkalk)	4.0	0.004	0.0035	0.01	0.05	0.9	406.09	7.14	40.61
Botovskaya Cave, Siberia, Russia, Lower Ordovician limestones	23.0	0.067	0.104	0.11	1.37	4.5	201.75	7.62	58.51
Estremera Cave, Madrid, Spain, Neogene gypsum	3.5	0.008	0.064	0.06	0.71	18.3	59.32	9.04	13.56
Optimistychna Cave, W. Ukraine, Neogene gypsum	188.0	0.26	0.52	1.48	26.03	2.8	127.03	2.00	17.57
Ozerna Cave, W. Ukraine, Neogene gypsum	111.0	0.33	0.665	0.74	13.2	6.0	150.00	5.04	44.59
Mlynki Cave, W. Ukraine, Neogene gypsum	24.0	0.047	0.08	0.17	2.38	3.3	141.18	3.36	27.65
Kristalna Cave, W. Ukraine, Neogene gypsum	22.0	0.038	0.11	0.13	1.82	5.0	169.23	6.04	29.23
Slavka Cave, W. Ukraine, Neogene gypsum	9.0	0.019	0.034	0.07	0.98	3.7	139.14	3.47	29.05
Verteba Cave, W. Ukraine, Neogene gypsum	7.8	0.023	0.047	0.07	0.66	6.0	117.82	12.00	34.74
Atlantida Cave, W. Ukraine, Neogene gypsum	2.52	0.0045	0.0114	0.02	0.29	4.5	168.00	4.00	30.00
Ugryn Cave, W. Ukraine, Neogene gypsum	2.12	0.004	0.008	0.01	0.14	3.8	176.67	5.71	33.33
Jubilejna Cave, W. Ukraine, Neogene gypsum	1.5	0.002	0.0035	0.01	0.08	2.3	277.78	4.00	37.04

Komsomol'ska Cave, W. Ukraine, Neogene gypsum	1.24	0.0017	0.0026	0.01	0.07	2.1	177.14	3.00	24.29
Dzhurinska Cave, W. Ukraine, Neogene gypsum	1.13	0.0016	0.0027	0.01	0.12	2.4	125.56	2.00	17.78
Zoloushka Cave, W. Ukraine, Neogene gypsum	89.5	0.305	0.712	0.63	18.93	8.0	142.06	3.76	48.41
Bukovinka Cave, W. Ukraine, Neogene gypsum	2.4	0.0043	0.006	0.02	0.14	2.5	120.00	4.44	21.50
Gostry Govdy Cave, W. Ukraine, Neogene gypsum	2.0	0.0013	0.0033	0.01	0.07	1.7	270.27	4.00	17.57

Notes:

1. Length of caves in the table corresponds to portions mapped by the dates when they were taken for this analysis. Further exploration has increased the length of some caves.

2. In the columns "Cave porosity" and "Areal coverage", values in brackets for the first three caves are those obtained by Worthington (1999) using a "rectangular" method for delineation of cave fields.

3. Calculations were performed using basic cave measurements and maps obtained or derived from the following sources: Blue Spring Cave, Mammoth Cave and Friars Hole System: Worthington (1999), Worthington *et al.* (2000); Jewel Cave and Wind Cave: Mark Ohms, personal communication (2000); Knock Fell Caverns: Elliot (1994); Fuchslabyrinth Cave: Müller *et al.* (1994); Moestroff Cave: Massen (1997); Botovskaya Cave: Filippov (2000); Estremera Cave: Almendros and Anton Burgos (1983).

TABLE 2

Average characteristics of conduit patterns for unconfined and confined settings

Parameter	**Settings**			
	Unconfined	Confined		
		Whole set	Gypsum caves	Limestone caves
Passage density, km/km^2	16.6	167.3	157.4	191.9
Areal coverage, %	6.4	29.7	28.4	33.0
Cave porosity, %	0.4	5.0	4.8	5.5

active fissures will be exploited in speleogenesis. The density of passages in the resultant network depends on the structural prerequisites. Variations in characteristics of fissure networks, resulting from the particular geological/tectonic conditions, can account for the above differences. It should be stressed that even though maze caves are the typical result of hypogenic transverse speleogenesis, they cannot form if the structural prerequisites are not favorable. For instance, on the other extreme of structurally-dependent hypogenic cave patterns are single fissure-like passages blind-terminated at both ends or rarely-intersecting fissure passages as encountered by mines in many regions such as in the Prichernomorsky artesian basin of Ukraine (Figure 31).

The second reason lies in the different speleogenetic history during the late artesian and post-artesian stages (transitional to unconfined conditions). Some caves or their parts may experience more intense growth than others during the transition from confined to unconfined settings, if they are favorably positioned relative to discharge points or zones. During the post-artesian stage, substantial volume in caves can be added due to horizontal notching during stillstands of the water table.

This study supports the conclusion drawn by Klimchouk (2003a) that any generalization of hydrogeology of karst aquifers, as well as approaches to practical hydrogeological issues in karst regions, should take into account the different nature and characteristics of conduit porosity and permeability that evolve in confined and unconfined settings.

5. Some implications of the hypogenic transverse speleogenesis concept

Despite obvious advances made during the last few decades, karst and cave science remains of limited significance and appreciation in such applied fields as formation of mineral deposits, hydrocarbon prospecting, groundwater management in artesian basins, mining, geological engineering, etc. One of the main reasons is that the predominantly epigenic karst paradigm, and respective concepts and knowledge of epigenic, unconfined karst, learned by industry geologists from general geology, groundwater hydrology and karst textbooks, was inappropriately and largely unsuccessfully applied to solving practical problems related to the quite distinct domain of hypogenic, confined, karst. The conceptual framework suggested in this book places hypogenic karst in the systematized context and hierarchical structure of basinal groundwater flow (in the sense of Tóth, 1999), and highlights the powerful role of speleogenesis in the organization of regional flow systems, a consequence of its unique capacity to dramatically alter the primary porosity and permeability of soluble formations. This framework suggests that karstified zones and their function in basinal groundwater systems are predictable. The new refined concept of hypogenic speleogenesis has broad implications in applied fields and promises to make karst and cave expertise more highly-valued by practicing hydrogeologists, mining engineers, and economic geology and mineral resource industries. A detailed discussion of all possible implications is far beyond the scope of this book, but below are a few particularly instructive examples and references given to illustrate the above contentions.

5.1 Variability in aquifer characteristics and behavior resulting from unconfined and confined speleogenesis

The specific mechanisms of ascending hypogene speleogenesis, discussed in Chapter 3, are responsible for the peculiar features of conduit porosity that develop in soluble formations under confined settings. This gives rise to characteristic distinctions between karst systems that develop in unconfined and confined karst aquifers. Huntoon (2000) provided an illustrative comparison of features found in unconfined and confined aquifers in Arizona, USA. The summary that follows is based on the discussions in previous sections (see also Klimchouk, 2003a) and the cited work of Huntoon.

Caves formed in *unconfined settings* tend to form highly localized linear or dendritic systems that account for high heterogeneity and extreme anisotropy of unconfined karst permeability. They receive more or less concentrated recharge from the immediately overlying or adjacent areas, with which they have genetic relations. Conduit systems are hierarchically organized to effectively concentrate and laterally transmit flow (and hence contaminants) in the downgradient direction. This organization is frequently cited to be similar to surface water drainage networks. Storage is commonly low in karst aquifers that evolved in unconfined settings, but almost all flow takes place through conduit systems (Worthington *et al.*, 2000). System responses to major storm events are characterized by flow-through hydraulics. Spring discharge from unconfined conduit systems tends to be flashy and highly variable.

The most common patterns for hypogenic caves formed in *confined settings* are 2-D or multi-story mazes in which conduits are densely packed, or complex 3-D systems. Hypogenic systems evolve to facilitate cross-formational hydraulic communication between common aquifers, or between laterally transmissive beds in a heterogeneous soluble formation, across the cave-forming zones. The latter commonly represent originally low-permeability units where vertical flow predominates. Caves receive either diffuse or localized recharge from the underlying aquifer or deeper parts of a succession. They do not have direct genetic relations with the overlying surface. This type of karstification commonly results in more isotropic conduit permeability pervasively distributed within highly karstified areas measuring up to several square kilometers but the actual pattern depends on the initial permeability structure. Localization of highly karstified areas depends on the distribution of head gradients in the multiple aquifer system (which is partly guided by erosional topography), heterogeneities in initial permeability of various beds in the system and on the nature and distribution of permeability in a feeding aquifer (source of cave-forming fluids). Although being vertically and laterally integrated throughout conduit clusters, confined conduit systems do not transmit flow laterally for long distances relative to the regional scale. White (1988) fittingly compared the organization of artesian maze systems with swamp hydrology.

Huntoon (2000) noticed that well-developed artesian karst porosity and storage in karst aquifers behave similarly to their counterparts in porous media, with the distinction that the "pores" are very large. Ubiquitous conduit porosity that develops through areas of transverse speleogenesis accounts for rather high aquifer storage. Discharge of artesian karst springs is commonly very steady, being moderated by high karstic storage developed in soluble units and by the hydraulic capacity of a whole artesian system.

5.2 The role of hypogenic speleogenesis in the formation of mineral deposits

The last two decades have seen rapidly growing recognition of the significance of fluid migration and groundwater flow systems in the genesis of mineral deposits; important reviews include Baskov (1987), Sharp and Kyle (1988) and Tóth (1988, 1999). A recent overview on the role of speleogenesis is provided by Lowe (2000). This section refines and reinforces some key aspects from the perspective of the new hypogene karst concept presented in this paper, and refers to some particularly instructive examples.

Sedimentary basins around the world that contain soluble carbonate and sulfate formations often host major epigenetic and stratabound deposits of metals (lead, zinc, barium, fluorine, copper, uranium, etc.) and sulfur, which appear to be associated with discharge segments of regional groundwater flow systems. The association of many such deposits with deep-seated karst features and high-permeability karstified zones was widely noted in the relevant literature that discussed their origin, geology, and hydrogeology. However, an important point was commonly missed, hindering more adequate understanding of mineral deposition. In contrast to the common views that karst porosity simply hosts mineral deposits, the refined concept of hypogene karst suggests that processes of deep-seated karstification and the formation of mineral deposits are dynamically linked (Klimchouk, 2000a). Mineral deposition not only fills or lines cavities or karst breccia zones, using them as spaces or guiding discontinuities, but it occurs because speleogenesis alters the regional flow system to converge at certain localities and creates necessary transitory reactive and depositional environments and geochemical thresholds.

Another aspect of the same problem is that, in interpreting the paleohydrogeology of mineral deposits associated with groundwater flow systems, high karst porosity is commonly taken as a given property of the hydrostratigraphic framework, assuming that "karst was always there" (either as paleokarst or with no consideration of its origin at all) to converge flow and/or host mineralization. Genetic and paleohydrogeology models for karst-related deposits almost always imply a hydrostratigraphic framework with highly permeable karstified aquifers and intervening non-karstic beds of low to moderate permeability. This is contrary to the common "initial" hydrostratigraphic framework that preceded hypogene speleogenesis.

The result of these misconceptions is the lack of recognition of the true genetic relationships between speleogenesis and ore formation, and some unresolved issues in models of ore genesis. This situation is mainly due to the fact, discussed in the introduction, that karst and cave science itself has so far failed to appreciate the significance of hypogenic/deep-seated speleogenesis, and to integrate the emerging relevant conceptual framework, mechanisms and methodology into the general karst paradigm. Karst and cave scientists had not yet offered hydrogeologists and ore and petroleum geologists the appropriate conceptual and terminological arsenal.

The place of hypogenic speleogenesis within a basinal flow domain is discussed in Section 2 and shown in Figure 1. It is regularly associated with discharge segments of regional or intermediate flow systems. However, the arguments of this paper and substantial evidence worldwide strongly suggest that this association is largely because speleogenesis creates these discharge segments, and makes them recognizable at the regional scale. Hence, in basins containing soluble formations, the primary result of speleogenesis is converging groundwater flow to zones where ascending cross-formational communication is greatly enhanced by speleogenesis, a condition commonly seen as the most important for flow-induced accumulation of transported mineral matter. In this way, speleogenesis facilitates the interaction of waters of contrasting chemistries and different geochemical environments. Thus, it often creates geochemical thresholds or transitional environments favorable to precipitation and accumulation of mineral ores, such as sulfide metals, sulfur, certain types of uranium deposits, etc.

Sulfur deposits

Large sulfur deposits are associated with gypsum and/or anhydrite and formed by the oxidation of H_2S generated by reduction of dissolved sulfates in the presence of hydrocarbons. The overall process results in epigenetic replacement of sulfate rocks by calcite and sulfur ores. Although general geochemical conditions and processes for the formation of epigenetic sulfur deposits are well established (Feely and Kulp, 1957; Ivanov, 1964; Yushkin, 1969; Ruckmick *et al.*, 1979; Kirkland and Evans, 1980; Machel, 1992; Kushnir, 1988), the genetic models for many deposits are still debatable, mainly in aspects of their paleohydrogeology. Proper flow models are crucial for understanding ore genesis since it is the appropriate conditions in a flow system that allow particular geochemical processes to operate and produce massive mineral accumulations. To generate large sulfur deposits, such hydrogeological systems must accommodate sources for dissolved sulfate and hydrocarbons, their interaction in an anaerobic environment to reduce sulfates to H_2S, and proper conditions for sulfide to oxidize to native sulfur at the boundary between the reducing and oxidizing environments.

Intense karstification is ubiquitously reported for virtually all epigenetic[2] sulfur deposits. Klimchouk (1997c) generalized that karstification is the intrinsically accompanying process for the formation of epigenetic sulfur because it supplies the dissolved sulfates needed for large-scale sulfate reduction. In turn, sulfate reduction serves to maintain the dissolutional capacity of groundwaters with respect to gypsum and anhydrite. Even more importantly, speleogenesis opens up cross-formational hydraulic communication paths and establishes flow patterns that provide the spatial and temporal framework within which the processes of the sulfur cycle take place. In this way, it controls geochemical environments and the migration of reactants and reaction products between them.

[2] Note that the term *epigenetic* (not *epigenic*!) is used here in the connotation of changes in the mineral content of a rock because of outside influence, occurring later than deposition of the host rock.

Epigenetic sulfur deposits form within multi-story confined aquifer systems. Most of them are associated with areas where the upper confining sequence is considerably scoured by erosion, *e.g.* within fluvial valleys or paleovalleys, which induces upward discharge in gravity-driven flow systems and transverse speleogenesis in sulfate beds. In mixed systems, where there is a prolific aquifer beneath a sulfate sequence, transverse speleogenesis can be supported or enhanced by the buoyancy component. Three regional examples below, from western Ukraine, northern Iraq and the Delaware Basin in the USA, are particularly illustrative of the role of speleogenesis in the formation of sulfur deposits.

Western Ukraine. The Miocene gypsum sequence is exposed along the southwestern margin of the eastern European Platform, in the transition zone between the platform and the Carpathian Foredeep. Gypsum extends from northwest to southeast for 340 km in a belt ranging from several kilometers to 40-80 km wide. It is the main component of the Miocene evaporite formation that girdles the Carpathian folded region to the northeast, from the Nida River basin in Poland across western Ukraine and Moldova to the Tazleu River basin in Romania. The Miocene succession comprises deposits of Badenian and Sarmatian age. The Lower Badenian unit, beneath the gypsum, includes carbonaceous, argillaceous and sandy beds (10-90 m thick), which comprise the main regional aquifer. The Middle Badenian gypsum sequence is up to 40 m in thickness and overlain by the Ratynsky evaporitic (a few meters thick) and epigenetic (up to 30 m thick) limestone. The latter variety has low $\delta^{13}C$ signatures ranging from -32 to -65‰, a diagnostic feature for bioepigenetic calcite recognized in major sulfur deposits around the world. This calcite contains sulfur ore in the deposits and locally replaces the gypsum entirely. The Ratynsky limestone and the lower parts of the overlying Kosovsky Formation comprise the upper (supra-gypsum) aquifer, overlain by the upper confining clays and marls of the Kosovsky Formation.

The sulfogenic province in the western Ukraine lies mainly within the confined zone of the Miocene aquifer system, which is recharged on the northeast where the confining sequence is eroded and the lower aquifer is exposed at higher elevations. The confined flow zone extends to the southwest, where it is dammed by the tectonic boundary with the Carpathian Foredeep. Discharge occurs throughout the confined flow zone via tectonic faults or karst breakdown structures. In the adjacent interior parts of the platform, the Miocene aquifer system is presently unconfined due to intense Plio-Pleistocene uplifts and deep erosional entrenchment. Extensive maze caves are known there from the same gypsum sequence, five of which are the longest gypsum caves in the world. They are shown to be the foremost examples of artesian transverse speleogenesis, being formed by dispersed recharge from the sub-gypsum aquifer (Klimchouk, 1990, 1996c, 2000b).

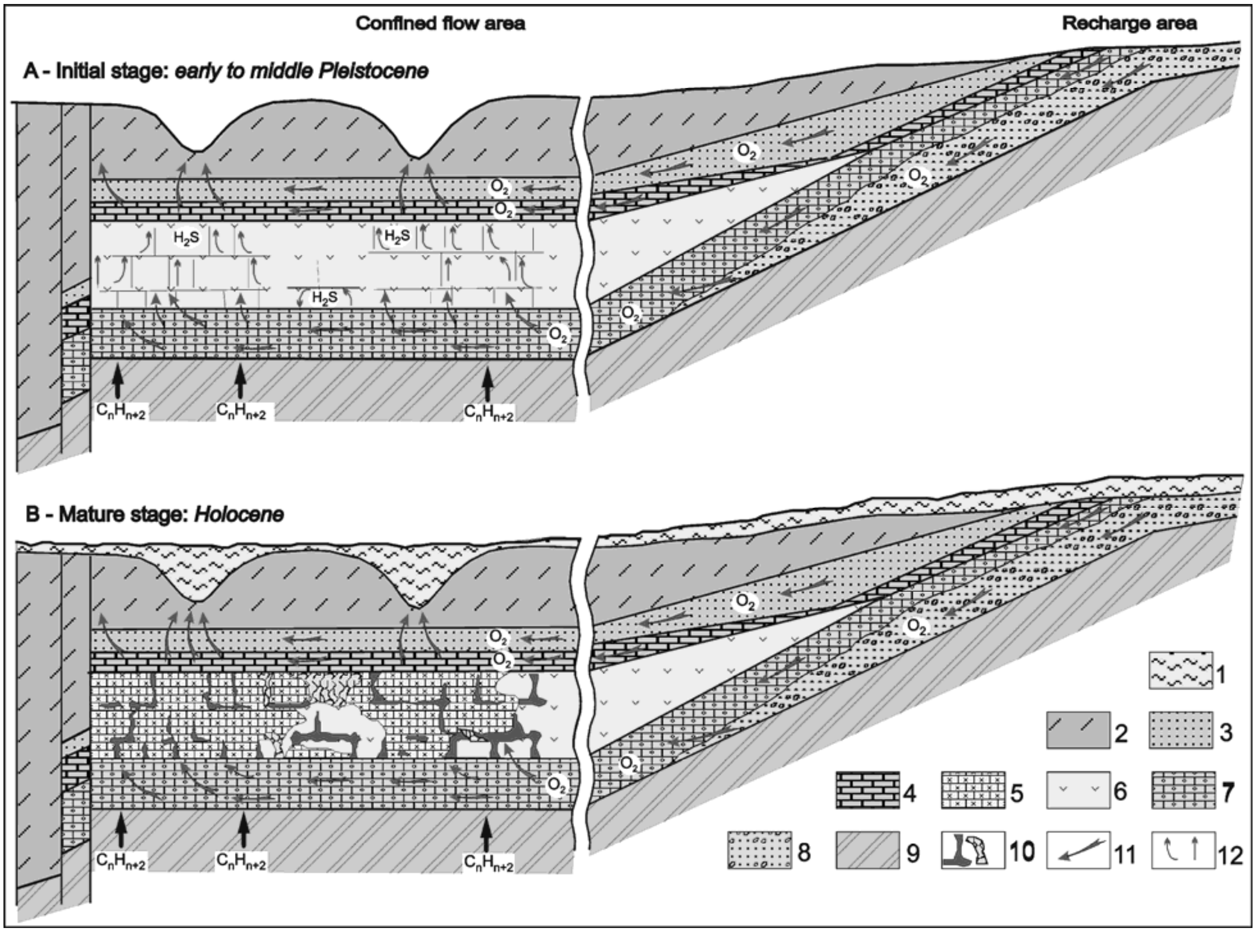

Figure 57. Conceptual model of speleogenetic origin of sulfur deposits in the western Ukraine (from Klimchouk (1997c). Key: *Quaternary sediments*: 1 = sands and loams. *Upper Badenian sediments*: 2 = clays and marls; 3 = sandstones; 4 = Ratynsky limestones; 5 = epigenetic sulfur-bearing and barren limestones; 6 = gypsum and anhydrite. *Lower Badenian sediments*: 7 = lithothamnion limestones; 8 = sands and sandstones. *Upper Cretaceous sediments*: 9 = marls and argillaceous limestones. 10 = dissolutional cavities; 11 = flow patterns in main aquifers; 12 = flow patterns through karst systems.

The long debates on the origin of sulfur deposits of the fore-Carpathian region had been hindered by an inadequate underlying hydrogeological model that treated the gypsum bed as an aquitard separating the aquifers in the Miocene system. Despite common awareness of the widespread occurrence of karst features in the sulfur deposits, both in the gypsum and limestone, the flow-forming role and structure of karst systems was not recognized in these models. Klimchouk (1997c) provided a comprehensive synthesis of regional geological, hydrogeological and karstological data to demonstrate that transverse speleogenesis in the gypsum played a fundamental role in the origin of sulfur deposits, by creating extensive high permeability clusters in the gypsum and by providing a favorable interposition of geochemical environments. Speleogenesis in gypsum created the necessary pattern of migration of reactants and reaction products between them to form bioepigenetic calcite and sulfur at the top of the gypsum. The speleogenetic model of the origin of sulfur deposits in the fore-Carpathians is shown in Figure 57. Aerobic conditions within the lower aquifer favored microbially-mediated transformation of methane to simple organic compounds that can be readily utilized by sulfate-reducing bacteria. Water from the lower aquifer ascended through the gypsum, forming aerially pervasive, although clustered, cave systems, with the upper part of the gypsum being the arena for intense sulfate reduction and gypsum-calcite replacement under anaerobic conditions. In the upper aquifer, the ascending H_2S was oxidized by O_2-bearing waters that came laterally through the upper aquifer, and vertically where steady buoyant plumes of water from the lower aquifer were established through the mature cave systems ("punctures" of oxidized waters through an otherwise reducing milieu). Paleohydrogeological analysis suggests that the most favorable timing for sulfur-ore formation was from the early to middle Pleistocene, although in some areas the same process is operative even today.

Northern Iraq. A karst-related origin is demonstrated for the main sulfur deposits of northern Iraq (Jassim *et al.*, 1999). The brief description that follows is derived from the cited work. This sulfogenic province is associated with the Middle Miocene Fatha Formation, which contains gypsum and/or anhydrite interbedded with carbonates, marls and claystones. The Fatha Formation contains aquifers in its carbonate beds and overlies the major aquifer in the oil-bearing Lower Miocene carbonates of the Euphrates-Jeribe Formation (Figure 58). The main deposits (Lazzaga, Mishrag and Fatha) are located along the course of the Tigris River, which partly incised into the Miocene sequence, created a regional piezometric low and induced upward discharge from the underlying confined system. Sulfur mineralization is mostly restricted to the lower member of the Fatha Formation. Isotopic signatures of sulfur are consistent with the microbial formation of the source H_2S, and calcite that replaces gypsum inherits a ^{13}C-depleted isotopic composition from hydrocarbons.

The model for the sulfur origin suggested in Jassim *et al.* (1999; Figure 58) invokes sulfur accumulation in cavities, dissolved in a gypsum bed that averages 10 m thick and is sandwiched between carbonate beds conducting lateral groundwater flow. Mineralization concentrates in zones where rising hydrocarbon-bearing waters from the Lower Miocene carbonates mix with lateral and downward influxes of oxygen-bearing water. According to this model, the cavity zone experiences alternating reducing/oxidizing conditions in response to the fluctuating rainfall, allowing for alternating reduction of dissolved sulfate and oxidation of H_2S to accumulate epigenetic calcite and sulfur. The model does not specify a speleogenetic style for the formation of "a cavity" depicted in Figure 58, but another work of Jassim *et al.* (1997) interprets gypsum karst in the region in conventional terms of "descending" surface-derived recharge and unconfined systems. To fit with the stratabound occurrence of the sulfur ore, the pattern of ore-hosting cave porosity should be laterally pervasive. This can be produced in the given hydrostratigraphic conditions only through confined transverse speleogenesis. The hydrostratigraphic arrangement described above seems very favorable to supporting confined speleogenesis and the generation of maze caves in the gypsum bed. In this case, the model for the origin of sulfur deposits in Northern Iraq will be largely like the above described model for the western Ukraine, suggesting the critical role of speleogenesis.

Delaware Basin, West Texas. Bioepigenetic sulfur deposits in the Delaware Basin are associated with the thick evaporitic Castile, Salado and Rustler formations of Upper Permian (Ochoan) age (see Figure 45 for location and stratigraphy). The evaporitic sequence conformably lies on the Lamar Limestone member of the hydrocarbon-bearing aquiferous Bell Canyon Formation, which is

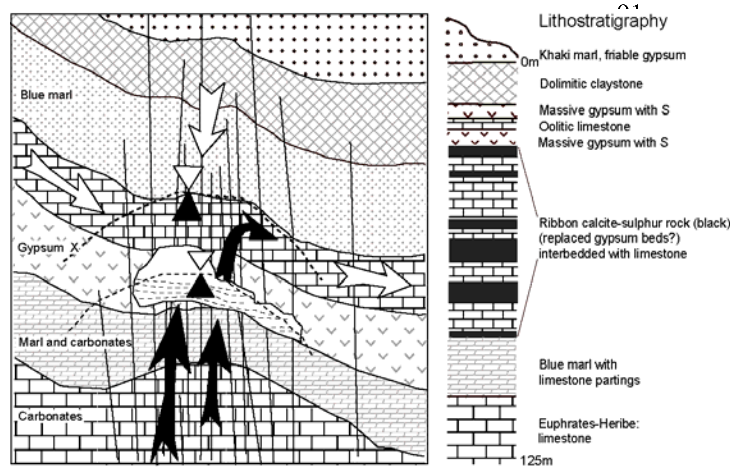

Figure 58. The conceptual genetic model for sulfur deposits of northern Iraq (from Jassim *et al.*, 1999). Black arrows indicate rising flow and white arrows indicate influx of oxygen-bearing water.

composed of limestones, sandstones and marls, a basinal equivalent of the Capitan Formation. The Castile Formation is 200 to 600 m thick, has a basal limestone member, but is mainly composed of anhydrite and gypsum, with minor halite interbeds largely removed by dissolution and substituted by homolithic breccia. The Salado Formation varies in thickness from 30 m at the western edge to 760 m in the center of the basin and is composed of sulfates and halite. The upper part of the Salado Formation contains most of the known sulfur deposits in the region. The Rustler Formation, 30-200 m in thickness, includes alternating siltstone, limestone, gypsum and dolomite. Above the Permian sequence lies the Dewey Lake Formation, composed of mudstone and siltstone. Predominantly clastic cretaceous sediments overly the Permian rocks through part of the basin.

All major works on the geology and origin of sulfur deposits in the Delaware basin underscore their close relationship with karst features, particularly with cross-formational features that allowed penetration of hydrocarbons with upward flow across the thick evaporites (*e.g.* Anderson and Kirkland, 1980; Kirkland and Evans, 1980; Miller, 1992; Wallace and Crawford, 1992). Of particular relevance are those vertical karst structures that extend from the basal limestone/sandstone aquifer. These include masses of epigenetic calcite (either barren or sulfur-bearing) called "buttes" or "castiles," which form exhumed hills on the eroded surface of the evaporites, and heterolithic breccia chimneys (pipes). Anderson and Kirkland (1980) recognized that density-driven convection was the important mechanism for deep dissolution beneath and across the evaporites, with fresh water from the basal aquifers rising into the evaporites and developing cross-formational cavities, and brines returning to the aquifer to ultimately outflow through it. At some point, such cavities collapse to form breccia chimneys, but they persist to act as high permeability structures that provide access for hydrocarbons to reducing zones and sulfates. Such breccia structures are heterolithic, being formed at different times from the Triassic-Jurassic through the present, based on the composition of heterolithic material and distribution in the basin (Wallace and Crawford, 1992). Epigenetic calcite bodies and sulfur deposits have been emplaced along cavities and breccia structures since the late Cenozoic Basin and Range extensional tectonism (Kirkland and Evans, 1980; Miller, 1992; Wallace and Crawford, 1992). Influx of oxygenated waters to interact with H_2S to form native sulfur occurred through shallow subsurface carbonate beds in otherwise evaporitic sequences (such as those in the Rustler), as well as through various disruptions of evaporites, lateral ramifications of the hypogenic karst structures beneath some barriers, and epigenic karst features induced by the development of hypogene features in the deeper zones. the Culberson ore body is an example of how most of the sulfur accumulations occur at the top of the Salado Formation, immediately beneath the vertically heterogeneous Rustler Formation. Some deposits, however, formed in the lower section of the Castile, immediately above the Bell Canyon aquifer. The type example is the Pokorny deposit (Klemmick, 1992), where the formation of sulfur-bearing calcite bodies was apparently guided by contact-type buoyancy-driven speleogenesis.

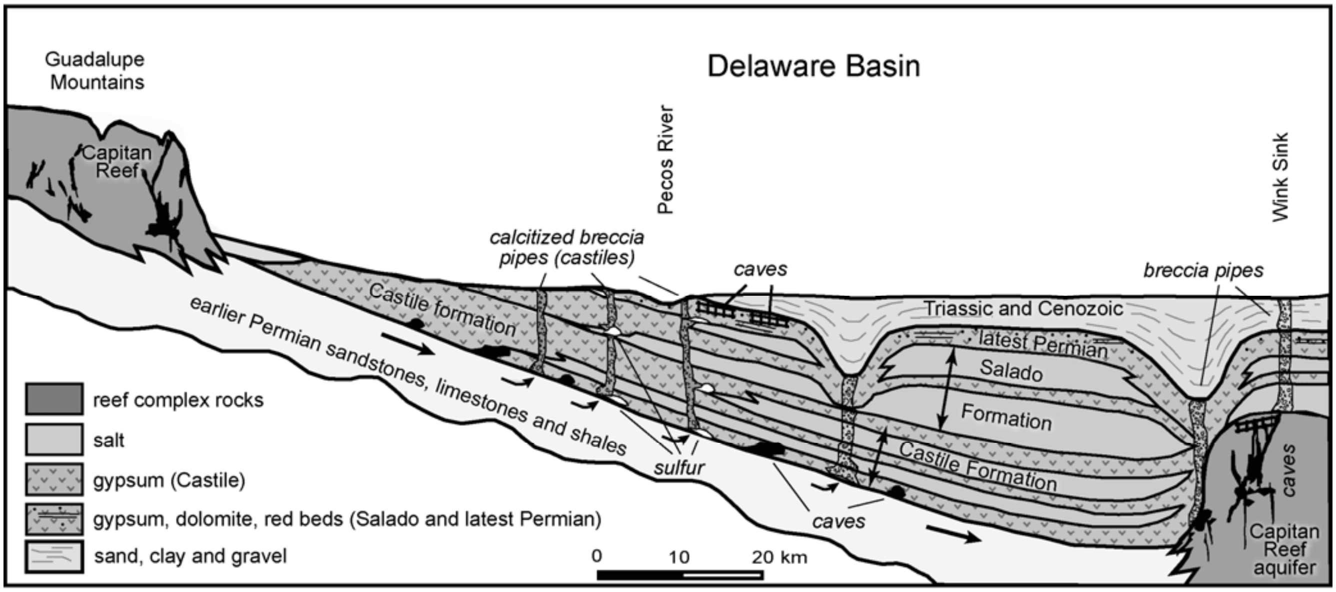

Figure 59. Diagrammatic representation of hypogenic karst features in the Delaware Basin and adjacent reef structures, New Mexico and west Texas, USA. Adapted from Martinez et al. (1998) for hypogenic features.

MVT lead-zinc deposits

MVT (Mississippi Valley Type) carbonate-hosted ore deposits are considered by various researchers to be the result of the mobilization, transport, and accumulation of metal ions by regional groundwater flow (Baskov, 1987; Garven *et al.* 1999; Tóth, 1999). Despite this general understanding, geologists continue to debate the mechanisms of fluid flow and chemical theories for ore deposition. The relevance of hypogene speleogenesis to the origin of MVT ore deposits has been recognized by some workers (*e.g.* Ford, 1986; Ghazban, *et al.*, 1991; Hill, 1996) but there is much broader potential here. To illustrate this, below is a summary of the hydrogeological characterization (based on Garven *et al.*, 1999) of the world's most important lead-zinc ore district, located in southeast Missouri, USA.

The sulfide ore districts in the Mississippi Valley region occur in Upper Cambrian and Lower Ordovician dolomite strata that blanket the Precambrian rocks on the Ozark Dome. Deposits are concentrated in a dolomitic reef facies of the Cambrian Bonnetere Formation, with ore-mineralization patterns controlled by pinchouts of the underlying Lamotte Sandstone (against the Precambrian granite) and collapse brecciation trends. It is believed that deep sulfate brines were topography-driven mostly northward from the source (the Arkoma foreland and underlying basement), with focusing of flow, heat and chemical mass within the carbonate formations. The general interpretation is that ore formation was concentrated on the Ozark Dome because of regional groundwater discharge, aquifer pinchouts, and favorable conditions for geochemical deposition related to permeability, cooling, and fluid mixing. In the context of the hypogene karst concept presented in this book, it can be easily seen that *these conditions are associated with confined transverse speleogenesis and are largely created by it.*

At the regional scale, nearly horizontal flow mostly occurs in the basal aquifer (Lamotte and Bonneterre Formations), which is confined by the thin Davis Shale. The hydrostratigraphy is shown in part 3 of Figure 60. The main deposits are aerially associated with a high-permeability lens, the Viburnum Trend, which is 20 km wide and about 100 km long. It is situated over the basement arch, but where the topographic elevations are the lowest. The Viburnum Trend affects regional flow patterns, creates a large discharge zone and focuses flow upon it, and induces a hydrothermal anomaly. Ore mineralization patterns at the deposit-scale are controlled by sandstone pinchouts, karstic channels and breccia zones (which are also karstic channels in this case).

It is beyond the scope of this paper to offer an elaborate speleogenetic/ore origin conceptual model for MVT deposits of the Ozark Dome, but the hydrogeologic conditions described above seem to be generally favorable for hypogene speleogenesis, which could be a major factor in the formation of ores:

- In the basal aquifer system regional lateral flow of brines occurred mainly through sandstone units, with unkarstified carbonate strata serving as intervening beds.
- The Ozark Dome area and local pinchouts of the basal sandstone within it were favorable sites for transverse speleogenesis to commence through the overlying carbonate Bonnetere Formation, due to the combined ascending potential of both the regional topography-driven flow system and local thermal anomalies induced by the direct rise of hot fluids from fractured Precambrian basement into the carbonates.
- Multiple dissolution mechanisms for speleogenesis could operate; various mixing effects (particularly invoked in these settings), dissolution due to increased calcite solubility in cooling hydrothermal paths, sulfuric acid, dedolomitization, etc. (see Section 3.6); this should be the subject of a separate region- and deposit-specific analysis.
- Transverse speleogenesis changed vertical permeability and opened migration paths across carbonate units and the confining bed (through fracturing and collapsing in response to growing cave porosity below), enhancing flow and regional discharge and inducing various reactions at geochemical thresholds that commonly occur along cross-formational paths. Some of these thresholds could favor ore deposition in previously created karst porosity at some stages.
- Fluid migration, speleogenesis and ore deposition were transient processes, adapting to the regional dynamics of landscape evolution as well as to deposit-scale dynamics of porosity and permeability changes. A number of geochemical models for ore formation can be adapted to fit the above hydrogeologic/speleogenetic scheme.

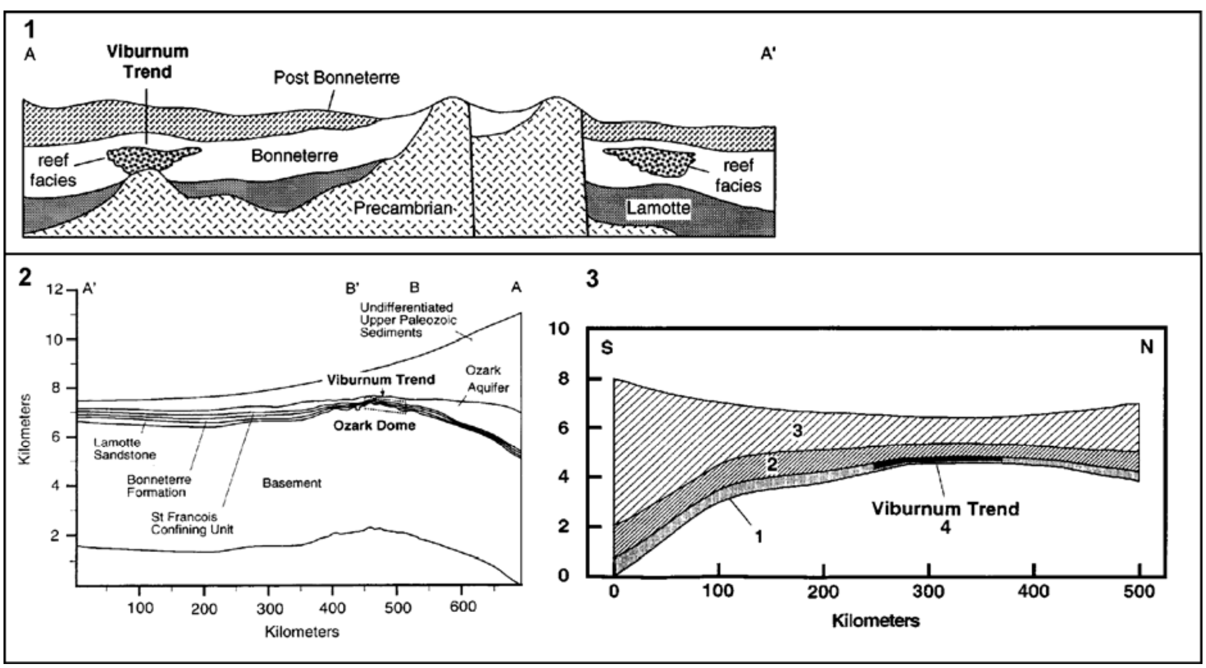

Figure 60. 1 = Geologic section A–A' in southeast Missouri, USA, showing geology of the Southeast Missouri Ore District. In the Viburnum Trend, several deposits formed in the reef facies of the Bonneterre Dolomite, and ore-bearing solutions appear to have migrated up from the Lamotte Sandstone. (from Kaiser *et al.* 1987). 2 = Lithostratigraphy of the Ozark Dome region; 3 = Hydrostratigraphy of the Ozark Dome region: 1 = the basal sandstone-carbonate Cambrian-Ordovician aquifer, with "high-permeability lenses" within it, rested on fractured Precambrian basement; 2 = less permeable Ordovician carbonates and shale; 3 = Permian shale (Adapted from Garven *et al.*, 1999). Note similarity of litho- and hydrostratigraphy on these sections with those of the speleogenesis model of Brod (1964) (see Figure 39)

5.3. Implications for petroleum geology and hydrogeology

As with ore deposits, the role of hypogenic transverse speleogenesis in converging flow and enhancing cross-formational hydraulic communication between stories in layered reservoirs can also be demonstrated for migration and concentration of hydrocarbons. As shown in Chapter 2, hypogenic speleogenesis is able to influence groundwater flow systems at the regional scale. The difference with respect to ore deposits is that entrapment of hydrocarbons, and the formation of oil and gas fields, is caused not by geochemical barriers but by stratigraphic and hydrodynamic barriers in overlying or laterally adjacent insoluble low-permeability units.

Many important deposits of hydrocarbons throughout the world are associated with karstified formations. An important issue in hydrocarbon exploration is characterization of karst porosity in production horizons in oil and gas fields. It is presently approached almost exclusively on the basis of general epigenic karst concepts, taken in the context of paleokarst. The most popular model is an island hydrology model implying speleogenesis at the freshwater/saltwater mixing zone beneath a limestone island. The concept of hypogenic transverse speleogenesis presented throughout this book opens new perspectives for interpreting karst features in oil and gas fields and applying karst and speleogenetic knowledge to industry needs.

The Permian Basin of west Texas and southeast New Mexico, USA, provides abundant examples of karst-related oil fields (Figure 61). Note that oil fields to the north and east of the Delaware Basin are aligned with buried sections of the Capitan Reef. In view of the confined hypogene speleogenesis model suggested for the Guadalupe Mountains (the presently exposed part of the reef; see Section 4.5 and Figure 49), it can be presumed that the buried reef section hosts hypogenic karst systems similar to those known in the Guadalupe Mountains, and that speleogenesis also affected the backreef facies (Yates, Seven Rivers and Queen Formations). On the Northwest Shelf and the Central Basin Platform, the Seven Rivers Formation serves as a leaky seal for the San Andres limestone, a host formation for many oil reservoirs. It now appears that both formations support intense hypogenic karst development. In the Guadalupe Mountains, much cave development occurred in the Seven Rivers and lower Yates (*e.g.* parts of Lechuguilla Cave and the McKittrick Hill caves). In the evaporitic facies of the Seven Rivers (north of Carlsbad), hypogenic speleogenesis in gypsum is exemplified by the study of Coffee Cave (Stafford et al., 2008). The series of huge sinkholes at Bottomless Lakes State Park on the eastern margin of the Roswell Artesian Basin (Land, 2003; 2006), and the Wink Sinks above the eastern sector of the Capitan Reef (Johnson et al., 2003), are formed by upward artesian flow from the San Andres and Capitan reef aquifers, further illustrating ongoing hypogenic speleogenesis under confined conditions. The oil fields in the south of the Central Basin Platform and the Midland Basin lie in the area where hypogenic transverse speleogenesis in the Cretaceous Edwards Group is exemplified by Amazing Maze Cave and Caverns of Sonora (see Section 4.5). This type of speleogenesis is probably dominant throughout the entire region. These are just brief references to demonstrate that the hypogenic transverse speleogenesis model is a sound alternative to the island speleogenesis model (paleokarst) when dealing with karst in the west Texas and southeastern New Mexico oil fields. Hill (1996) provided discussion of the relevance of sulfuric acid speleogenesis to petroleum deposits of this region.

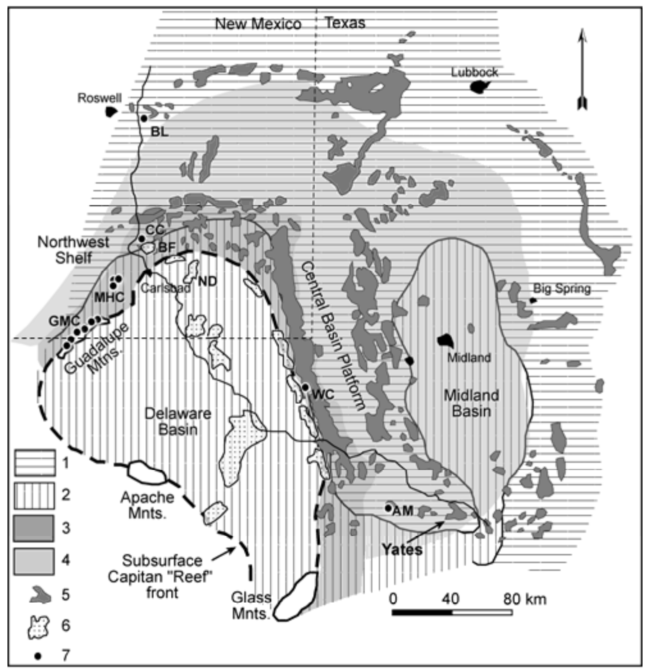

Figure 61. Distribution of oil and gas fields of west Texas and southeast New Mexico in relation to major features of regional geology and hypogenic karst. Compiled using data about regional geology features from Scholle *et al.* (2004) and oil and gas field distribution from Craig (1988) and Ward *et al.* (1986). Shading indicates distribution of Permian, lower Guadalupian depositional facies: 1 = San Andres limestone and dolomite; 2 = Lower Cherry Canyon & Brushy Canyon formations (basinal sandstone, siltstone & shale); 3 = Backreef dolomites and sandstones of Yates, Seven Rivers, & Queen formations; 4 = Evaporite facies. 5 = Oil and gas fields; 6 = Dissolution troughs in evaporites; 7 = Major features of hypogenic karst: BL = Bottomless Lakes; CC = Coffee Cave; BF = Burton Flat; ND = Nash Draw; MHC = McKittrick Hill caves; GMC = Guadalupe Mountains caves; WC = Wink Sink; AM = Amazing Maze Cave.

An instructive example of both a karstified oil reservoir and the classic approach to interpreting karst features in it, is the Yates oil field on the southeastern corner of the Central Basin Platform (Craig, 1988). The field has shown remarkable production characteristics and abundant evidence of intense karstification such as bit drops, sudden rushes of oil during drilling, extremely high flow rates, etc. In 92 of 400 wells from this field, used for the analysis by Craig (1988), unfilled cavities were documented. This gives a 23% probability of hitting cavities, which can be compared with an average 29.7% of areal coverage shown by typical hypogenic maze caves (see Tables 1 and 2). The above value of areal coverage, however, is calculated from the "cave fields," whereas the probability of wells hitting cavities for the Yates field is derived from the sample of wells not necessarily located within cave clusters. Distribution of wells that hit caves in the field (Figure 16.4 in Craig, 1988) clearly shows a clustered pattern characteristic of hypogenic transverse speleogenesis.

Caves encountered in the Yates field average 0.9 m in height and range from 0.3 m to 6.4 m, which are typical of confined maze caves such as the presently relict Amazing Maze Cave, located in an adjacent area above the production horizon of the White and Backed oil fields. They are concentrated in the upper 15 m of the San Andres Formation, but also occur in several other stratigraphic intervals.

In accordance with established views, karst features in the Yates field were interpreted as late Permian paleokarst, and their spatial distribution was speculatively fitted to a model of a freshwater/saltwater mixing zone beneath a cluster of small limestone islands, which were hypothetically emergent in the Permian seas. It is the present author's contention that the model of ascending hypogenic speleogenesis is a more feasible alternative not only to the Yates field but for the majority of carbonate-hosted petroleum reservoirs in the Permian Basin region.

5.4. Implications for sinkhole hazard and site assessments

The sinkhole hazard problem is commonly approached from the perspective of surface investigations and studies of subsurface structures by oblique methods, *e.g.* geophysics, drilling, etc. Caves inherently lie at the core of the problem, but the potential for gaining deeper and more adequate understanding of sinkhole-forming processes from a speleological perspective remained largely unexploited, prompting Klimchouk and Lowe (2002) to draw attention to this possibility.

The speleogenetic approach to the problem is very promising. Clearly, the difference in cave porosity structures created by epigenic and hypogenic speleogenetic processes and respective groundwater flow systems points to potential peculiarities of sinkhole formation processes.

Klimchouk and Andrejchuk (2005) provided an instructive case study of sinkhole formation processes in intrastratal (entrenched, subjacent and deep-seated) gypsum karst of the western Ukraine, using extensive hypogenic maze caves to map and investigate breakdown structures at the cave level (Figure 62). Extrapolated density of breakdown structures (localities where cavities had collapsed and the caprock sediments penetrated into the cave unit) varies from a few hundred to over 5,000 features per square kilometer between different caves and morphologically distinct areas of large caves, although only a small proportion of these structures propagate through the overburden to cause sinkhole expression at the surface. It was found that, in contrast to the conventional wisdom, distribution of breakdown structures does not appreciably correlate with the size of the passages and rooms. The study has shown that breakdown is initiated mainly at specific speleogenetically or geologically "weakened" localities, which fall into a few distinct types. Most breakdowns that are potent enough to propagate through the overburden relate with the outlet cupolas/domepits that represent places where water had discharged out of a cave to the upper aquifer during the period of hypogenic transverse speleogenesis. This is because, by virtue of their origin and hydrogeological function within a hypogenic transverse system, such features had exploited the points of lowest integrity within the main bridging unit of the upper aquifer and the entire overburden. The study also gave an important insight into the mechanisms of breakdown propagation to the surface and demonstrated numerous potential implications of the speleogenetic approach for more adequate and efficient sinkhole hazard assessment in areas of hypogenic karst in stratified sequences.

Vertically extensive breccia pipes (or vertical breakdown structures) known from many regions of the world are related to yet another type of hypogenic transverse speleogenesis, where large cavities are formed due to buoyancy-driven upward dissolution at the base of evaporite formations, with fresh water rising from the basal aquifer and dense brine sinking and outflowing through the aquifer (Anderson and Kirkland, 1980; Kempe, 1996). Subsequent collapsing of cavities gives rise to the formation of breakdown structures (Figure 59). These propagate through upward-stoping maintained by active groundwater circulation, accompanied by dissolution and suffosion (Huntoon, 1996; Klimchouk and Andrejchuk, 1996). The developing structures drain any intercepted aquifers and serve as pathways facilitating and focusing vertical hydraulic communication across thick sequences. Outstanding examples are breccia pipes within the Phanerozoic sedimentary succession of the Grand Canyon region, Arizona, USA. Recent mapping of fossilized and epigenetically calcified breccia pipes ("castiles"), performed in the Gypsum Plain region of the Delaware Basin, USA, identified 1,020 features over an area of 1,800 km^2, which suggests an average density of 0.57 features/km^2 (K .Stafford, personal communication).

However, their distribution is clustered, with the density of castiles within clusters reaching 18 features/km^2. That hypogenic speleogenesis is an ongoing process throughout the region and capable of generating large sinkhole development is exemplified by the Bottomless Lakes series of sinkholes east of Roswell, New Mexico, or the Wink Sink collapse on the eastern margin of the Delaware Basin, Texas (see Figure 61 for their location).

Proper recognition of hypogenic transverse speleogenesis gives a new perspective to such important issues as assessment of sites of special concern, such as the Waste Isolation Pilot Plant (WIPP) near Carlsbad, New Mexico, or some nuclear power plants. The most karst-specific studies for the WIPP area addressed the issue from the perspective of epigenic karst concepts (Bachman, 1990; Hill, 2003b; Lorenz, 2006; Powers *et al.*, 2006), which leads to a misleading interpretation of observed features when dealing with hypogenic karst. Recent identification of the hypogenic origin for many caves in the Rustler and Seven Rivers Formations (Stafford et al., 2008) and interpretation of regional karst development in the context of hypogenic transverse speleogenesis calls for the need of a reassessment of karst hazard for the WIPP site. Another example of special concern is the Neckar nuclear power plant in Germany, situated in a river valley above a multi-aquifer system containing a sulfate bed, with ground deformation recorded in the immediate vicinity of the plant and a major 60 m-deep collapse that occurred in 1964 a few kilometers away. An additional site of concern is the Rovensky nuclear power plant in Ukraine, where old and recent collapse and subsidence features were recorded in the immediate vicinity, induced by hypogenic karst in the confined Cretaceous chalk aquifer, a part of the Volyno-Podol'sky artesian basin.

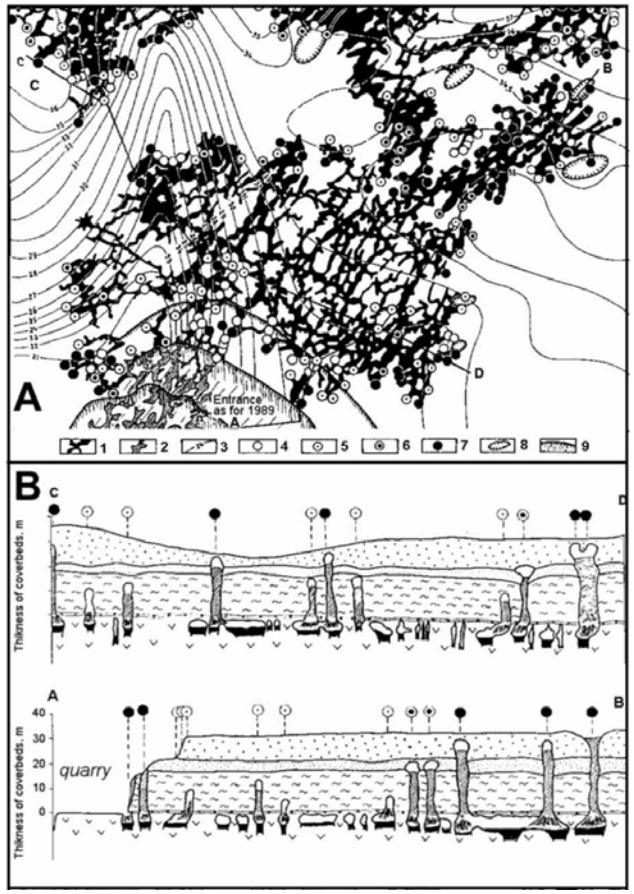

Figure 62. Distribution of breakdown structures in Zoloushka Cave in plan (A, map fragment) and profile (B). 1 = cave passages; 2 = passages destroyed by the quarry; 3 = isopachytes; 4 - 7 = breakdown structures with the breakout cavities positioned at various levels: 4 - at the bottom of the overlying aquifer, 5 - within the confining clays, 6 - within the sandy-gravel bed, the upper aquifer, 7 - within the loam bed; 8 = surface karst features; 9 = the quarry faces (from Klimchouk and Andrejchuk, 2005).

Epilogue

The major points and conclusions of this book are summarized in the abstract. This book does not pretend, and hence does not succeed, in covering all aspects of hypogenic karst and speleogenesis or in providing comprehensive regional overviews. Such an attempt would probably be premature because the conceptual framework of hypogenic speleogenesis is still newly established and poorly integrated into the main body of karst science. Instead, the main goal of this work was to consolidate the notion of hypogenic karst as one of the two major types of karst systems and to outline an approach that would help to see the forest behind the trees. This approach implies that speleogenesis should be viewed in the context of regional groundwater flow systems (not only of local systems that evolve through a soluble rock after its exposure), and their evolution in response to basinal processes, uplift, denudation and geomorphic development.

Various styles of hypogenic caves that were previously considered unrelated, specific either to certain lithologies (*e.g.* western Ukrainian giant gypsum mazes) or chemical mechanisms (*e.g.* sulfuric acid caves or hydrothermal caves) appear to share common hydrogeologic genetic backgrounds. They were formed by ascending transverse speleogenesis, which is responsible for the remarkable similarity of their most characteristic morphologic features. It is suggested that confined and semi-confined settings are the principal hydrogeologic environment for hypogenic speleogenesis, and that vertical heterogeneity in permeability is the principal control over hypogenic cave development. Evidence for this is overwhelming.

However, there is a general evolutionary trend for hypogenic karst systems to lose their confinement due to uplift and denudation and due to their own development. Confined hypogenic systems may experience substantial modification or be partially or largely overprinted under subsequent unconfined (vadose) stages, either by epigenic processes or continuing unconfined hypogenic processes, especially when H_2S dissolution mechanisms are involved. This means that in dealing with unconfined karst settings and epigenic caves, a possibility of inheritance from hypogenic cave development should not be overlooked or underestimated. It is likely that many caves, previously explained from the perspective of established epigenetic models, will be re-interpreted to more adequately account for such inheritance.

Hypogenic confined systems evolve to facilitate cross-formational hydraulic communication between common aquifers, or between laterally transmissive beds in heterogeneous soluble formations, across cave-forming zones. The notion of cross-formational hydraulic communication, quite well established in mainstream hydrogeology, was not properly realized in karst science. Hypogenic speleogenesis is essentially a cross-formational, transverse, phenomenon.

One of the main characteristics of hypogenic speleogenesis is the lack of genetic relationship with groundwater recharge from the overlying or immediately adjacent surface. It may not be manifest at the surface at all, receiving some expression only during later stages of uplift and denudation. But long before this expression occurred, and long before we got physical access to explore them, hypogenic caves were already there! And they are there[3], at some depth beneath a non-soluble confining cover, through vast areas normally not considered as karst, based on the traditional, largely geomorphological karst paradigm. There is abundant evidence of hypogenic caves, including those more than ten times greater than the largest cave chamber directly explored by humans, encountered by wells and mines at depths up to many hundreds of meters. Industry geologists and hydrogeologists deal with them routinely, but rarely karst scientists. However, those who deal with deep-seated karst features often fail to adequately understand them because the mainly epigenic models for karst and caves are readily available from geosciences texts. So, the common approach to deep-seated karst features is to put epigenic karst models into paleokarst wrapping. But this does not always help to effectively deal with karst-related issues, simply because they are often related to hypogenic karst, not to true paleo (epigenic) karst.

The refined conceptual framework of hypogenic speleogenesis has broad implications in applied fields and promises to make karst and cave expertise more highly-valued by practicing hydrogeology, geological engineering, economic geology, and mineral resource industries. Any generalization of the hydrogeology of karst aquifers, as well as approaches to practical issues and resource prospecting in karst regions, should take into account the different nature and characteristics of hypogenic and epigenic karst systems.

An appreciation for the wide occurrence of hypogenic karst systems, specific characteristics of their origin and development, and their scientific and practical importance, calls for revisiting and expanding the current predominantly epigenic paradigm of karst and cave science.

[3] This statement, however, implies not an anthropocentric definition of caves but the notion of "*a karst cave as an opening enlarged by dissolution to a diameter sufficient for 'breakthrough' kinetic rates to apply if the hydrodynamic setting will permit them. Normally, this means a conduit greater than 5-15 mm in diameter or width*," after Ford and Williams (2007).

Acknowledgements

Many individuals and institutions have helped the author in various ways to gain data, observations, experience and inspiration to develop ideas and concepts discussed through this book. I'd have to write another chapter to properly acknowledge their roles, but this would be a different genre that would go beyond an allowed volume and style of this book. I remember them all and I am grateful to everyone who happened to help me.

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In karst and cave science, it is particularly important to have first-hand knowledge of regions and features. Caves are not an easy subject for documentation and observations, and most of these science-related activities are based on original explorations and surveys done by cavers. My thanks go to many Ukrainian cavers who over the years explored the great gypsum mazes of the western Ukraine and assisted in studying them, and to many caving friends in other countries. Generalizations endeavored in this book would be impossible without people who were instrumental in various periods in arranging my field trips to some exemplary hypogenic caves discussed in this book: John Scheltens, Ron Kerbo and Jim Goodbar (USA), Dr. Paolo Forti (Italy), Dr. Jose Maria Calaforra and Sergio Garsia Dils (Spain), Dr. John Gunn and Dr. John Lamont-Black (UK), Ralph Müller, Wolfgang Pikart and Graham Hash (Germany), members of the Speleo-Club of Luxembourg, and many other caving friends.

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Front cover: A rising chain of cupolas in Caverns of Sonora, TX, USA (Photo by A. Klimchouk)
Back cover: A dome possibly leading to a higher storey of passages (an exploring caver climbing a rope provides a scale). See Plate 11 for a broader view. Echo Chamber in Lechuguilla Cave, NM, USA (Photo by S. Allison).

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