Karst Waters Institute

Special Publication 18

Hypogene Cave Morphologies

Edited by
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2014

Selected papers and abstracts of the symposium held February 2 through 7, 2014, San Salvador Island, Bahamas

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Cover caption: Precision radial cave passage survey in progress at Altar Cave, San Salvador, Bahamas.
Photo credit: Ira D. Sasowsky

HYPOGENE CAVE MORPHOLOGIES

San Salvador, Bahamas
February 2-7, 2014

CONFERENCE ORGANIZERS

John Mylroie
Alexander Klimchouk
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PREFACE

Welcome to San Salvador, the landing place of Christopher Columbus in 1492. That event changed the world - what changes in speleological thought will result from this meeting of minds in a very special karst environment? San Salvador is a small island on a small platform. Everything we see in terms of geology and karst had to happen in a constrained area in a very brief amount of geologic time. This compression in space and time provides a useful setting to explore how speleogenesis works.

The Hypogene Cave Morphologies conference is being held at the Gerace Research Centre (GRC) February 2-7, 2014. In keeping with the tradition of prior Karst Waters Institute meetings, the goal is creative interactions between engaged researchers. The main thematic activities of the conference are to examine and discuss the unique cave morphologies and speleogens associated with hypogene caves, from the scale of 100 km+ cave maps down to centimeter size wall rock shapes and forms. Hypogene caves can be argued to represent a laminar flow regime that is quite different from the turbulent flow found in epigenic stream caves coupled to surface hydrology. Can these morphologies be uniquely characterized to identify hypogene caves? What effect do these laminar flow regimes have on geochemical models of dissolution drive in hypogene settings? Do flank margin caves fall in the hypogene flow environment?

The conference begins with an optional visit to two very large but easily accessible flank margin caves on Eleuthera Island that have a broad suite of morphologies commonly associated with hypogene caves. The field trip will also visit two other sites that bring into question the separation of pseudokarst sea caves from coastal dissolution caves. The main part of the meeting is a field conference, with on-the-ground activities every day. Evenings and two half-days are devoted to sessions; with posters being displayed during the evening social hour. The first day field trip establishes the geologic setting of San Salvador, so that the caves and karst seen on the second day can be understood in the tight time and space constraints of the island. These constraints help illuminate possible mechanisms of cave formation in this environment. Lighthouse Cave, the largest flank margin cave on the island, is visited the third day, and the last day has a series of optional field trips in the afternoon.

About 30 invited participants from 13 countries, all of whom have demonstrated interest and expertise in hypogene caves, are participating. This volume presents their written contributions as abstracts and papers.

Conference Organizers

John Mylroie
Alexander Klimchouk
Ira D. Sasowsky

KEYNOTE PAPERS

HYPOGENE SPELEOGENESIS – DISCUSSION OF DEFINITIONS

Y.V. Dublyansky[1]

Existing definitions of the term *hypogene karst* (hypogene speleogenesis) are not always consistent with the established meaning of the term hypogene in the Earth Sciences. They are commonly biased either toward geochemical or toward hydrogeological aspects of the phenomenon. It is proposed that hypogene karst is defined on the basis of the two properties: predominance of the deep-seated sources of aggressiveness of karst water, independent of the environment at the overlying or immediately adjacent surface; and recharge of soluble formation from below, independent of recharge from overlying or immediately adjacent surface.

THE TERM HYPOGENE IN EARTH SCIENCES

The term *hypogene* was first introduced in geological literature by Sir Charles Lyell, in an attempt to remove ambiguity in rock designation "primary" and "secondary", customary at the time. Lyell (1833) postulated that "*the hypogene rocks can only originate at great depths in the regions of subterranean heat.*" Subsequently, the term was re-introduced by Ransome (1913) for ore deposits and minerals; again, to confront the "primary-secondary" ambiguity (mineral, ore, mineralization, etc.). He also proposed the complementary term supergene for minerals "*deposited by generally downward moving and initially cold solutions*" (p. 153).

Presently, *hypogene* is rarely used to designate rocks (i.e., sensu Lyell), but most commonly refers to processes, minerals, ores, deposits, solutions, and the like (sensu Ransome). This can be illustrated by one of the earliest definitions of Bastin and Laney (1918): "*Solutions coming from great depth within the earth and having a general upward flow will therefore be termed hypogene and the work they accomplished hypogene mineralization.*"

Similar understanding is captured in modern reference sources, e.g. "*Hypogene is a term used only in the adjective form for any geological process genetically connected with deeper parts of the Earth's crust or for mineral, rock and ore formed beneath the surface of the Earth. It is an opposite term to supergene. … Hypogene (= endogene) processes include tectonic, magmatic, metamorphic and hydrothermal processes, as well as the formation of various ore deposits. … Hypogene ore deposits are formed under the influence of deep-seated endogene (= hypogene) geologic processes by the action of ascending solutions. Such solutions and conditions of ore formation are also called hypogene*." (Springer Reference)

DEFINITION OF HYPOGENE SPELEOGENESIS

Finding a good definition for a natural phenomenon is always a balancing act involving search for the most appropriate mix of the exactness and the generality. Because the term hypogene is so widely used in the Earth Sciences, it is important also that its meaning in karst studies is consistent with the conventional one.

As the subject of hypogene karst gained prominence over the last decades, two approaches toward defining it have emerged. These approaches can conventionally be classed as geochemical and hydrogeological ones.

In a geochemical approach, the emphasis is on the geochemical mechanisms, processes of dissolution, and sources of aggressiveness that lead to creation of karst cavities. For example, Worthington and Ford (1995), following Ford and Williams (1989), define hypogene caves as those formed "*by processes involving sulfate, sulfide, and/or thermal waters*" (p. 9). A more-general definition was proposed by Palmer (2000) who defined hypogenic caves as those "*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.*"

The hydrogeological approach was introduced by Klimchouk (2007), who suggested that "t*he 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*" (p. 6), and "*The systematic transport and distribution mechanism capable of producing and maintaining the required disequilibrium conditions is the groundwater flow system … 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*" (p. 8).

Within the hydrogeological approach, hypogene karst is defined through its place and position in hydrogeological system, most commonly on the basis of its relationships with the recharge area at the surface. For example, Ford (2006) defines hypogenic speleogenesis 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 over-*

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lying or immediately adjacent surface" and Klimchouk (2007) postulated that the **only** main attribute of the hypogenic karst is its "*lack of genetic relationship with the groundwater recharge from the overlying surface*" (p. 6).

DISCUSSION

Karst is an undeniably hydrogenic process, and a hydrogeological approach in karst studies is therefore perfectly justified. Many concepts of the "mainstream" hydrogeology, such as the concept of the hierarchical character of hydrogeological systems for example are directly applicable to karst studies (Tóth, 1999; Klimchouk, 2007). There is, however, an important distinction, or rather, difference in focus, between the "classic" hydrogeology and the karst studies.

The former concerns itself primarily with the movement of water. Processes of water-rock interaction, although appreciated, do not occupy a central position in it. Aqueous geochemistry, the main branch of hydrogeology dealing with chemical effects of water-rock interaction, studies to what extent this interaction affects the water, but does not typically consider how it affects the rock. In contrast, in karst studies the effect of water-rock interaction on the rock (primarily, dissolution) is the sin qua non. In other words, without considering water-rock interaction (dissolution), hydrogeology remains hydrogeology. In contrast, taking out the water-rock interaction (dissolution) effectively annihilates the very subject of karst studies.

Because of this, it is my opinion that the balanced definition of the hypogene karst must integrate both approaches. Accordingly, the two defining properties of the hypogene karst are:

- Predominance of the deep-seated sources of aggressiveness of karst water, independent of the environment at the overlying or immediately adjacent surface; and
- Recharge of soluble formation from below, independent of recharge from overlying or immediately adjacent surface.

It is to be noted that, the notion of the "source of aggressiveness" is treated here in the most generic sense. For hypogene karst in carbonate rocks this role can be played by deep-seated acids. For rocks, which dissolve by simple dissociation, the very presence of unsaturated water can become a source of aggressiveness. The cooling of the upwelling water can be a source of aggressiveness for rocks having retrograde solubility (e.g., gypsum and calcite). And so on.

Importantly, the demonstrable lack of the deep-seated sources of aggressiveness in waters which enter the soluble formation from below, places speleogenesis in the "grey area," where its attribution to hypogene category becomes equivocal. A solution can be to designate karst (speleogenesis) which possesses both defining properties hypogene s*ensu stricto* (*s.s.*) and that which possesses only one of the defining properties – hypogene karst *sensu lato* (*s.l.*).

COMPLEMENTARY TERMS

The relevant complementary terms widely used in general geology and ore geology to define rocks, processes, minerals, mineral deposits are: *hypogene – supergene* (var. *hypergene*) and *endogene – exogene*. These "traditional" pairs of terms are semantically consistent. The first pair discriminates definienda in categories of relative position "above – below"

hypo- (from Greek) - under, below;

super- (from Latin) - above; over; beyond;

hyper- (from Greek) – over; above; cognate with Latin *super*-.

The second pair uses for discrimination categories "inside-outside"

endo- (from Greek) – within;

exo- (from Greek) – outside.

In karst lexicon, the pair *hypogene – epigene* (karst, cave, speleogenesis) is now entrenched. In contrast to the cases discussed above, the prefix *epi*- used to construct the term epigene has multiple meanings, some of which are not complementary to *hypo*-. Alternately, it can mean: on, upon, above, over (as in epicenter; epifluorescence); after (as in epilogue); in addition to (as in epiphenomenon); and near, close to (as in epicalyx). The term *epigenesis* is widely used in Earth Sciences, employing *epi*- in the second, temporal sense, along with a complementary *syngenesis*. Again, the latter pair is semantically consistent, referring to temporal categories "together, with" and "after".

A separate issue, specific to karst studies, is the similarity of the terms *epigene karst* and *epikarst*. The latter is defined as "*the uppermost weathered zone of carbonate rocks with substantially enhanced and more homogeneously distributed porosity and permeability, as compared to the bulk rock mass below*" (Klimchouk, 2004). Both terms employ suffix epi- in its spatial sense, with the distinction that the former refers to the "*origin from above*" while the latter refers to the "*location above.*" The epikarst is a part of the epigenic karst.

Summarizing, the term *epigene* (or epigenic) karst (speleogenesis) is not the most appropriate to define the opposite of the *hypogene* karst. The more appropriate antonyms would be *hypergene* or *supergene*.

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THE METHODOLOGICAL STRENGTH OF THE HYDROGEOLOGICAL APPROACH TO DISTINGUISHING HYPOGENE SPELEOGENESIS

Alexander Klimchouk[1]

Defined in the most general way, hypogene speleogenesis is the origin of caves in which the cave-forming agency comes from depth, in contrast to epigene speleogenesis in which the cave-forming agency (meteoric recharge and its inherent or soil-derived aggressiveness) originates at the surface. A more specific definition should rely on attributes of the cave-forming agency which are most suitable and efficient for discrimination between epigene and hypogene origin of caves.

Relying on the determination of a source of the aggressiveness in distinguishing hypogene speleogenesis is the legitimate approach but it is not a methodologically sound and practically efficient one.

The hydrogeological approach and the reference to upwelling groundwater circulation in the definition of hypogene speleogenesis provide a theoretically and methodologically sound basis not only for identifying the type of speleogenesis, but also for spatial and temporal prognosis of hypogene speleogenesis.

INTRODUCTION: APPROACHES TO DEFINE HYPOGENE SPELEOGENESIS

Advancements in karst and cave science during the past 20-30 years have led to the growing recognition of the possibility, wide occurrence, and practical importance, of conduit porosity development in deep-seated conditions, without direct influence of near-surface factors. Hypogene speleogenesis has become one of the hottest topics in karst and cave science, and the subject draws the increasing attention of other branches of geosciences, as well as of practitioners, particularly in the mineral and hydrocarbon resources exploration and in geological engineering.

However, there are some differences in approaches on how to define hypogene speleogenesis. Palmer (2000a) defined hypogenic caves as those *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 approach emphasizes the source of aggressiveness, and it is termed here "geochemical".

With the "hydrogeological" approach, hypogene speleogenesis is defined as the formation of solution-enlarged permeability structures by water that recharges the cavernous zone from below, independent of recharge from the overlying or immediately adjacent surface, which places an emphasis on the groundwater circulation system (GCS) (Ford, 2006; Klimchouk, 2007, 2013). This definition directly indicates that hypogene speleogenesis develops by upwelling flow, whereas the geochemical definition does not require this.

Are these approaches contradictory? In my opinion, they are not, although they impose somewhat different perspectives on the subject. Because of this, they determine different sets of speleogenetic environments and different samples of caves to be considered of the hypogene origin. This is a source of confusion and uncertainty that needs to be eliminated.

The aggressiveness of upwelling flow, in most cases, has been produced at depth, independent of near surface processes, and this is what constitutes the large common body of objects, outlined by both definitions. However, in some cases groundwater can keep the original undersaturation (i.e. aggressiveness) from distant recharge areas while moving deep underground along non-soluble aquifers in an artesian system and then ascending in discharge areas through soluble rocks. Hence, it cannot be said that the aggressiveness has been produced at depth in these cases but the aggressive water enters the cave-forming zone from below. This situation is especially common of hypogene speleogenesis in deep-seated evaporites. Moreover, dissolution of evaporates is "*independent of surface or soil CO_2 or other near surface acid sources*," as well as some other dissolution mechanisms such as dissolution in mixed carbonate/sulfate strata (dedolomitization). In other cases, the aggressiveness can be produced at depth as a result of mixing of two non-aggressive waters of contrasting chemistries along the interface while none of the waters is upwelling. Examples are freshwater lenses over saline water in homogenous eogenetic carbonates in island flank-margin environments, or downward infiltration water mixing with phreatic water at the water table.

Based on the geochemical approach, Palmer (2007) places the artesian transverse cave development in evaporites into the realm of epigene speleogenesis, whereas cave development due to mixing along hydrochemical interfaces in unconfined aquifers is placed into the hypogenic category. Within the hydrogeological approach advocated by the present author, the classifying of speleogenesis in these respective environments is the opposite.

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GENERAL DISCUSSION

Moving and aggressive groundwater is the principal cave-forming agency. Speleogenesis (karstic) is a mass-transfer process, which critically depends on both, the aggressiveness of groundwater and its circulation (movement). This equally applies to epigene and hypogene speleogenesis. Defined in the most general way, hypogene speleogenesis is *the origin of caves in which the cave-forming agency comes from depth*, in contrast to epigene speleogenesis in which the cave-forming agency (meteoric recharge and its inherent or soil-derived aggressiveness) is originated at the surface.

The question of a more specific definition is not about which of the attributes of groundwater, the aggressiveness or circulation, is more important for speleogenesis. The emphasis on a GCS in the hydrogeological approach does not mean that the importance of the aggressiveness (dissolution) is neglected, as dissolution is inherently implied as the integral part of the definitions of *karst* and *speleogenesis* (karstic). The question is about which of these attributes is most suitable and efficient to discriminate between epigene (hypergene) and hypogene speleogenesis.

It should be remembered that a definition of a natural phenomenon not only classifies a set of respective objects by referring to their most essential common attributes, but it also determines to a large extent methodologies to identify and study the phenomenon. The latter aspect is particularly important for hypogene speleogenesis, as in most cases we deal with relict caves decoupled from the cave-forming environments. The identification of a cave origin relies on our ability to discern characteristics of the cave-forming environments and processes from studying their indirect indications preserved after the environments have changed and the processes have ceased. It relies, therefore, on which of the attributes of the cave-forming agency are referred to in a definition, and on how they are represented in our study objects.

AGGRESSIVENESS

Aggressiveness is an attribute of groundwater that corresponds to a chemical potential for mobilization of a dissolved matter from the rock. It results from disequilibrium in the water-rock system that is created by the groundwater circulation.

It has to be noted that it is the aggressiveness that is an attribute of moving groundwater, but not the opposite. This attribute is the transitional one. It can originate and cease in a given segment of the circulation system, and migrate through the latter with changing intensity and pattern of circulation. Also, the nature of the aggressiveness can change during the evolution of a GCS, and dissolution can proceed through different chemical mechanisms, which are much more varied in hypogene speleogenesis that in epigene. For a given hypogene cave system, different mechanisms may operate either simultaneously or in a sequence, and it is often difficult to impossible to discern which of them has contributed most to speleogenesis. We normally have limited indications at our disposal to judge the dissolution mechanisms that operated in the formation of now relict caves, or about where the aggressiveness has been produced at the time of speleogenesis (although its origin at depth below the cave-forming zone is commonly implied). Mineralogical indications are useful but they rarely tell us about characteristics of principal stages of speleogenesis *per se*. Isotopic and geochemical traces of water-rock interactions in host rocks can be a strong evidence of hypogene speleogenesis (see Spoetl and Dublyansky, in this volume), but they are not always present or preserved.

Another fundamental question is whether principal characteristics of caves (their patterns, morphology, functioning, and distribution) are determined by differences in a source of the aggressiveness. In other words, does using a source of the aggressiveness as the main criteria for defining types of speleogenesis give us a useful tool to discern meaningful families of speleogenetic objects? The answer is *yes* perhaps only for epikarstic porosity. It is apparently *no* for karstic porosity that forms in phreatic conditions or in water table settings.

The corollary from the above discussion is that relying on the determination of a source of the aggressiveness in distinguishing hypogene speleogenesis is the legitimate approach, but that it is not a methodologically sound and practically efficient one.

GROUNDWATER CIRCULATION

Circulation (movement) is an inherent attribute of groundwater. Both the spatial distribution and efficiency of dissolution are controlled by intensity and a pattern (vector) of the groundwater circulation. The above-mentioned major characteristics of caves (cave morphology in particular) are determined not only by where the aggressiveness is produced relative to the surface, but also by how dissolution effects are distributed. The latter is dictated largely by the hydrodynamic characteristics of a GCS. Hence, it is the GCS that has to be a primary consideration for discrimination between the types of speleogenesis.

The primacy of the hydrogeological settings of a karst aquifer in determining the cave patterns has been demonstrated by Palmer (1991, 2000a). This is because hydrodynamics imposes a primary control on the location and development of conduits, as well as on shaping their morphology. The location and distribution of void-conduit systems and their patterns are determined by the overall pattern of GCS, the position of soluble rocks in relation to it, and the recharge and discharge conditions. Hence, the differences in origin and the development mechanisms of karstic void-conduit systems (types of speleogenesis) are determined largely by hydrodynamic peculiarities of GCS.

At the very broadest scale, two types of GCS are recognized according to hydrodynamics: 1) confined (to a varying degree) stratal and fissure-vein systems, and 2) predominantly unconfined near-surface systems. Accordingly, two fundamental types of speleogenesis can be distinguished: 1) *hypogene* in confined systems by upwelling circulation across soluble rocks from distant or separated by insoluble layers external or internal recharge sources, and 2) *epigene* (hypergene) in hydraulically open settings by downward and lateral circulation from overlying or immediately adjacent recharge surfaces. The differences in hydrodynamics between the respective GCS impose major distinctions in the mechanisms of these types of speleogenesis.

HYDRODYNAMIC CONTROL ON SPELEOGENESIS

In unconfined near-surface settings, discharge through conduits is controlled by two conditions (Palmer, 1991): 1) the hydraulic capacity of conduits (hydraulic control) or, 2) the amount of available recharge from the surface (catchment control). During the early stages of speleogenesis, the positive feedback between discharge and the growth of initial conduits causes their competitive and selective development. With the accelerated growth of conduits after breakthrough, they quickly reach dimensions at which the fixed head condition at the recharge boundary cannot be supported any longer so that the initial hydraulic control switches to the catchment control. Further development of conduits is characterized by their competition for the surface recharge, which determines the further increasing selectiveness in the process and close genetic relationship between epigene speleogenesis and karst geomorphogenesis. Hence, in epigene speleogenesis the positive feedback between discharge and the growth of conduits strongly operates not only during the early speleogenetic stages (Palmer, 1991; Dreybrodt, Gabrowsek, Romanov, 2005), but also during the mature stage.

In confined and semi-confined settings, where flow is directed transversely across layers and formations, both recharge and discharge of conduits occurs through adjacent insoluble beds (or segments in fissure-vein systems) with a relatively conservative permeability. Discharge in the whole GCS is controlled by the least permeable elements in the cross-section. Before the onset of speleogenesis, such elements are commonly represented by beds of soluble rocks, and discharge through early conduits in them is controlled by their hydraulic capacity. When transverse conduits reach the breakthrough condition, their further growth does not accelerate dramatically, as it occurs in epigene speleogenesis, because the control over discharge switches to the permeability of adjacent or more distant insoluble beds. The switch to the external conservative control over discharge in hypogene speleogenesis subdues the positive feedback loop and the speleogenetic competitiveness. This difference in speleogenetic mechanisms (epigene and hypogene) is one of the fundamental causes of distinctions in structure and morphology between the respective void-conduit systems. Another fundamental cause is the difference in the vector of groundwater circulation, which is explored below.

UPWELLING CIRCULATION

The hydrogeological definition of hypogene speleogenesis directly relates it with upwelling groundwater circulation. Even in relict systems, the past presence of the upwelling circulation can be recognized in most cases from the morphogenetic analysis of caves or/and from the paleohydrogeological analysis. The locations, in which upward flow is or was dominating, are 1) mappable from hydrogeological data, at least in basinal settings, and 2) predictable from regional hydrogeological analysis (for actual GCS) and paleohydrogeological/paleogeodynamic analysis (for past GCS). Hence, the reference to this attribute in the definition provides a methodologically feasible basis not only for identifying the type of speleogenesis, but also for spatial and temporal prognosis of hypogene speleogenesis.

The immanent link of hypogene speleogenesis with upwelling flow is suggested by vast empirical evidence and justified theoretically. The upward circulation dominates in the lower stories of the geohydrosphere because of the presence of internal recharge sources, the ultimate openness of circulation systems at the upper hydraulic boundary, and overall decrease of pressure toward the surface. During the geostatic and endogenous stages of the basin development the upwelling circulation may encompass the most of the sedimentary cover. The upward branch is also an important component of circulation in the upper part of the geohydrosphere, in the domain of the hydrostatic (meteoric) regime, where the overall circulation is determined by the balance between the downward and upward branches.

Modern hydrogeology acknowledges an immense importance of the vertical hydraulic communication (leakage) across low-permeability layers separating aquifers in meteoric GCS (Mjatiev, 1947; Hantush and Jacob, 1955; Shestopalov, 1981; Toth, 1995). Such communication in the meteoric regime is directed downward beneath highlands and upward below topographic depressions. The upward flow below topographic depressions is traced up to depths of 1 – 1.5 km, and it is generally more intense and localized than the downward flow beneath highlands at compatible depths (Shestopalov, 1981).

The most fundamental reason why hypogene speleogenesis is linked with upward circulation, but not with the downward circulation, lays in the speleogenetic mechanism. As noted above, the overall vertical permeability of the heterogeneous cross-section is determined by the least permeable intervals. In the areas of upward circulation, initial speleogenesis in soluble beds increases their permeability. This, in turn, steepens the hydraulic gradient across the upper insoluble confining unit and hence, the overall discharge in the system (Fig. 1). This re-activates the positive feedback loop and stimulates further development of transverse conduit. The gradient and discharge further increase with continued erosional entrenchment to the upper confining unit. In contrast, in the areas of more diffuse downward circulation, hydraulic resistance to flow increases with depth. Moreover, possibilities for internal discharge are limited. This prevents an

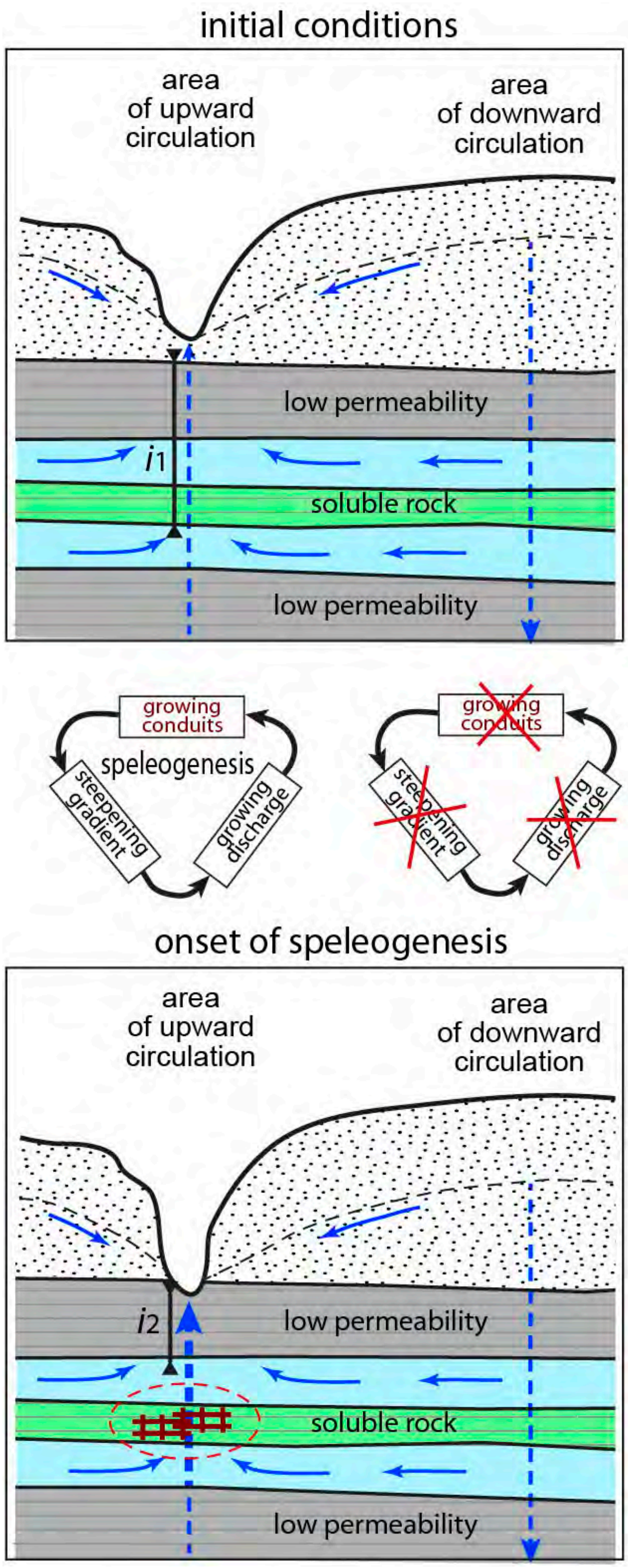

Figure 1. A conceptual illustration of speleogenetic potentials in the areas of upward and downward circulation in a layered aquifer system.

increase in the circulation intensity and inhibits the mechanism of speleogenesis (Fig. 1). Similar arguments can be used for vertical flow in a cross-formational fracture-vein structure that crosses rocks of variable lithologies including soluble ones.

Another important peculiarity of confined (semiconfined) hydrogeological environments is their low fluid dynamics as compared to unconfined settings, which favors to the natural convection circulation at the conduit (void) scale. Its effects are commonly well expressed in cave morphology (Klimchouk, 2007, 2009). The dissolution effects of natural convection are linked, again, with the upwelling limbs of convection cells but not with the plunging ones.

GROUNDWATER REGIMES

The overall circulation regime of a groundwater system is determined by the nature and magnitude of fluid pressure and by the degree of hydrodynamic confinement of the GCS. Different types of circulation regimes are distinguished in the literature.

In subsiding basins the dominant flow drive in progressively buried strata is compaction due to the increasing load, which causes expulsion of the pore waters from the sediments. This is the *geostatic regime*, also termed expulsion regime (in the Western literature), or elision or *exfiltration regime* (in the East European literature). Flow in such systems is directed upward, and on the regional scale - from areas of greatest subsidence to the margins of basins. The expulsion GCS are unlikely to play a role in hypogene speleogenesis.

With still deeper burial and further rise of temperature and lithostatic load, the *thermobaric regime* develops in which the fluid pressures are caused by the thermal expansion of water or by the release of water by mineral dehydration in a low-permeability environment. The *compression regime* can be generated by tectonic strain in the vicinity of collision and uplift areas. In the East European literature these two regimes are commonly combined into the *endogenous regime*, which also includes localized intrusions of fluids into the sedimentary cover from the lower crust and the upper mantle. The upward migration of endogenous fluids is considered to be the main cause for hydrogeochemical inversions and a phenomenon of the column-like desalinization observed in the lower parts of the sedimentary cover in many basins (Ezhov, 1978; Lukin, 2004). Upward flow dominates in the endogenous regime. The endogenous GCS, characterized by high temperatures and pressures, are believed to be very potent to support hypogene speleogenesis in a variety of rocks (Dublyansky, 2000; Andreychouk et al., 2009; Klimchouk, 2012).

The *free convection regime* may develop in some settings, especially in the vicinity of hydrothermal anomalies and in strata comprising evaporites, driven by density differences. The upwelling limbs of convection GCS are capable of supporting hypogene speleogenesis, especially in evaporites.

Following uplift and continental exposure, the *hydrostatic regime* evolves, driven by topography differences. It is also termed the *meteoric regime* (in the Western literature) and the *infiltration regime* (in the East European literature). With the continuing exposure and geomorphic development, meteoric waters increasingly flush out the formation waters from basins so that the hydrostatic regime substitutes the geostatic regime in the upper part of the crust, although the latter may still predominate in deep environments.

INTERACTION BETWEEN GROUNDWATER SYSTEMS OF DIFFERENT REGIMES

The meteoric regime is perched on ubiquitously upwelling waters of the geostatic and endogenous regime, commonly overpressured (the fluid pressure exceeds the normal hydrostatic one). Zones of interaction between GCS of different regimes, either crosscutting or lateral, are particularly favorable for hypogene speleogenesis in carbonate rocks because mixing of waters differing in CO_2 or H_2S content or salinity generates aggressiveness. Hypogene speleogenesis is commonly a part of mixed flow systems, where topography-driven flow interacts with the deeper compaction- or density-driven regimes, or with plumes of endogenous waters. The nature and the geometry of the transition between the different regimes are controlled by respective fluid potentials and geological heterogeneities, especially sedimentary windows and conductive faults. The lateral boundaries may be blurred, but they are more distinct when they coincide with low-permeability strata of regional extent. With the onsets of uplift and denudation in the course of geological evolution, the deeper strata migrate relatively upward through these boundaries, and the nature and geometry of the transition adjusts to the structure of uplifting strata and changing potentials of the interacting regimes.

HYPOGENE SPELEOGENESIS FROM THE PERSPECTIVES OF REGIONAL HYDROGEOLOGICAL ANALYSIS

As noted above, the association with upwelling circulation suggests the possibility of discerning regularities of development and distribution of hypogene speleogenesis from the perspectives of a regional hydrogeological analysis.

In basinal settings, the pattern of the meteoric circulation is controlled by the basin geometry and relief, by geological inhomogeneities that determine permeability distribution, and by interaction with deeper GCS of the geostatic and endogenous regimes, which may pierce through the domain of the hydrostatic regime. In mature artesian basins of the cratonic type, settings favorable for upward flow and hypogene speleogenesis, are as follows:

1) marginal areas of discharge of groundwaters of the 2nd hydrogeological story (HG-story);
2) zones of topography-controlled upward circulation within the internal basin area (at the 1st and, in places, at the 2nd HG-stories);
3) crests of anticlinal folds or uplifted tectonic blocs within the internal basin area where the upper regional aquitard is thinned or partially breached; and
4) linear-local zones of deep-rooted cross-formational faults conducting upward flow from internal deep sources across the upper HG-stories.

Hydrodynamics in the 3rd and 4th HG-stories in the cratonic basins are dominated by upward circulation (geostatic or endogenous regimes) strongly controlled by cross-formational tectonic structures.

Specific circulation patterns develop in large Cenozoic carbonate platforms (the Florida-type), side-open to the ocean, where upward flow across stratified sequences in the coastal parts, driven by both topography-induced head gradients and density gradients, involves mixing with seawater. At deeper levels, the seawater can be drawn into a platform along permeable horizons and rise in the platform interior due to geothermal heating (Kohout's scheme), interacting with upper freshwater aquifers.

In young basins where the geostatic regime dominates, hypogene speleogenesis is favored at marginal discharge areas where circulation systems of different origins and regimes may interact. Examples are meteoric systems circulating from adjacent uplifted massifs, basinal fluids expelled from the basin's interiors, and endogenous fluids rising along deep-rooted faults.

The predictability of the distribution of areas of upwelling flow in tectonically deformed mountainous regions is significantly lower than in cratonic basins because of the complexity and variability of geological and structural conditions, relief, and a geodynamic history in such regions. Massifs in the folded regions are characterized by dominance of fracture-vein groundwater systems, although sequences of the upper structural story often host stratal aquifer systems. Upward flow and hypogene speleogenesis in massifs are tightly controlled by faults, especially by those at junctions between large tectonic structures and structural stories, and by the geodynamic evolution. Specific and very favorable settings for hypogene speleogenesis are found in regions of young volcanism and hydrothermal activity.

Hypogene speleogenesis may also occur in deep oceanic settings, especially in regions associated with plate boundaries and hot spots. An outstanding example, although interpreted differently, is represented by extensive fields of large-scale depressions in the Mio-Pliocene carbonate blanket at depth of 1500-2600 m in the volcanic Carnegie Ridge, located within the Galapagos hotspot in the Pacific Ocean, recently documented by high-resolution multibeam bathymetry (Michaud et al., 2005). The host carbonates do not contain shallow facies and have never been subaerially exposed, which excludes any epigene karstification.

THE ROLE OF CONFINEMENT

In discussing the origin of maze caves, many of which are believed to form under artesian conditions, Palmer (1991, 2000b) argued that slow groundwater flow near chemical equilibrium, typical of truly confined aquifers, is least likely to produce maze caves. He further stressed that "*True confinement by itself does not produce maze caves, and any association between confined groundwater flow and maze development is coincidental*" (Palmer, 2000b, p. 79). The problem of the origin of maze caves is beyond the scope of this paper; it is considered in details by Palmer (1975, 1991, 2000a, 2000b, 2007, 2011) and Klimchouk (2000, 2007, 2009). Here it is appropriate to clarify some misconceptions about confinement, with regard to hypogene speleogenesis.

Mylroie and Mylroie (2009) provide a lengthy discussion on whether confined flow is necessary to produce hypogene caves. They argue that the morphological features believed to be characteristic of hypogenic caves in the hydrogeological connotation of this term (Klimchouk, 2007, 2009) are not solely the result of confined hypogenic conditions, but also occur in eogenetic karst aquifers, in environments that have never been confined, and have never undergone burial or been moved out of the influence of meteoric diagenesis.

The present author agrees that true confinement by itself does not produce maze caves. It has to be noted that the term 'confined aquifer' is not used in modern hydrogeology in a sense of a true hydraulic isolation, so that "true confinement" simply does not exist. Although a certain degree of leakage was long accepted to occur even through aquicludes, it was during the last 40-50 years that the great role of transverse hydraulic communication across separating beds in basins has been acknowledged. The "classical" artesian paradigm, with its notions of confined flow through largely isolated aquifers, was replaced with the basin hydraulic paradigm, with its notions of a multiple aquifer system and significant cross-formational (across aquitards) communication between aquifers. The adoption of this paradigm to karst studies has eliminated the ground for the above-mentioned concern and opened a new perspective to the problem of speleogenesis in artesian settings (Klimchouk, 2000, 2007). The abovecited works show that the association between confined groundwater flow and hypogene cave development is not coincidental, and that artesian transverse speleogenesis is one of the most common variants of hypogene speleogenesis.

The question whether confinement by itself is a necessary condition for hypogene speleogenesis is somewhat misleading. The term "confined" refers to a hydrodynamic condition wherein 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 overlying confining unit, and this allows water to move up through available, more permeable, paths. Hence, at least a certain degree of hydrogeological confinement is a necessary condition for the forced ascending groundwater circulation to occur. It is the upwelling circulation, but not confinement by itself, which is considered in the hydrogeological approach to be the main condition for hypogene speleogenesis, although confinement is certainly the common characteristic of flow in saturated heterogeneous media.

Another misconception (Mylroie and Mylroie, 2009) is that confinement always implies that a carbonate sequence must be once buried and moved to the mesogenetic realm. In fact, it does not. Confined (pressurized) flow may occur in sequences of eogenetic limestones, as they commonly demonstrate distinct layered heterogeneity formed due to variations in depositional and post depositional processes. For instance, Budd and Vacher (2004) show that matrix permeability of young carbonates in the Upper Floridan Aquifer range over three orders of magnitude between different lithofacies. Cunningham et al. (2006) developed a high-resolution cyclostratigraphic model for the Plio-Pleistocene 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.

It was shown (Girinsky, 1947) that where a vertical head gradient exists between aquifers in a layered sequence, and if hydraulic conductivities in adjacent beds differs by at least two orders of magnitude, flow in high conductivity beds is predominantly lateral, but flow in the separating beds is predominantly vertical. The above data on heterogeneity of eogenetic carbonate sequences suggest that they may host confined (leaky) aquifer systems with a characteristic pattern of interaction that may include rising transverse flow components.

Although hypogene speleogenesis develops mainly in confined conditions, it is not limited to them. When hypogenic caves are shifted to the shallower, unconfined situation due to uplift and denudation but their further development continues to be driven by upwelling flow from deeper systems, this is still hypogene speleogenesis, although now partly unconfined. Unconfined hypogene development can be regarded as an extinction phase of hypogene speleogenesis in most cases. However, the cave development fed by upwelling recharge at the bottom of an unconfined aquifer in eogenetic carbonates can also be considered to be hypogenic.

HYPOGENE SPELEOGENESIS IN EOGENETIC CARBONATES IN COASTAL REGIONS

Cave development in eogenetic carbonates in coastal settings is

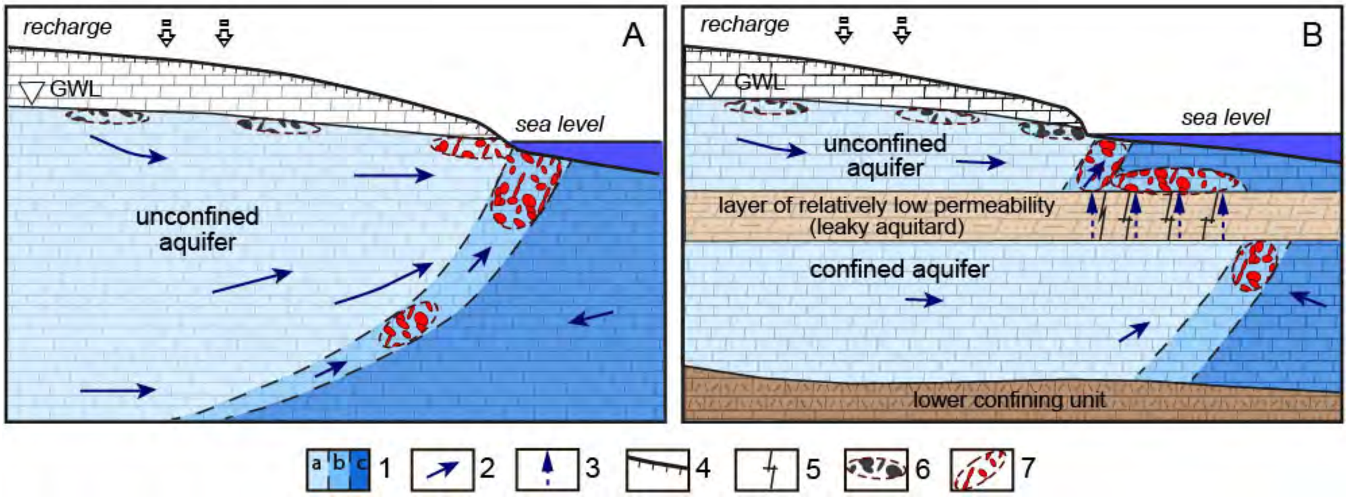

Figure 2. Speleogenesis in coastal areas: A = the standard flank-margin model for homogenous rocks (redrawn after Mylroye and Carew, 1995); B = an expanded model with elements of layered heterogeneity (the hydrogeological setting is borrowed from Barlow, 2003). Legend: 1 = groundwaters: a - fresh, b - brackish, c - saline (marine); 2 = flow directions; 3 = ascending leakage across the aquitard; 4 = epikarst; 5 = fractures or other conductive discontinuities across the aquitard; 6 = speleogenesis by mixing of vadose and phreatic freshwaters along the water table; 7 = speleogenesis by mixing of freshwater and marine water. Note that the speleogenesis by mixing of freshwater and marine water in cartoon B would be hypogenic speleogenesis according to the hydrogeological definition.

described by the flank margin model (Mylroie and Carew, 1995). Caves form as the result of mixing of freshwater and seawater at the bottom and especially at the distal margins of a floating freshwater lens. Because the aggressiveness is produced at depth within the bedrock mass, these caves are considered to be hypogenic within the geochemical approach (Mylroie and Carew, 1995; Palmer, 2007). As the standard model considers a floating Dupuit-Ghyben-Herzberg freshwater lens and cave development in unconfined phreatic conditions, the flank-margin caves were not regarded as hypogenic according to the hydrogeological approach.

Mylroie and Mylroie (2009) provide a number of illustrations showing a great deal of similarity between flank-margin caves and confined hypogenic caves formed by upwelling flow. They argue that the characteristic morphological features of flank-margin caves form due to slow flow conditions in the mixing zone that allow natural convection to extensively operate, and that the upwelling limbs of natural convection cells play a pronounced role in shaping the passage morphology. This is indeed a feasible explanation for the above-mentioned similarity. Flow in confined aquifers is also commonly slow, and the great role of natural convection circulation in shaping hypogene caves has been demonstrated and underscored (Klimchouk, 2000, 2007, 2009). It has to be noted that upwelling flow is in place in both cases.

A question remains for the case of flank-margin caves, however, as to whether their morphogenesis is solely due to the natural convection circulation self-developed along the freshwater/marine water interface in a homogenous rock, or whether it originates by upward leakage (recharge) of a freshwater aquifer from a layer of high hydraulic conductivity (a confined aquifer) below, across a separating layer of relatively low conductivity (an aquitard)? In such case, the caves would be classified to be properly hypogenic according to the hydrogeological approach. One could expect the presence not only of certain characteristic wall and ceiling bedrock features in such caves, but also the entire "*morphological suite of rising flow*" (Klimchouk, 2007, 2009), including feeders. This suite, but not separate features, was considered to be truly diagnostic for hypogene caves, as it unambiguously indicates upwelling circulation of the cave-forming fluid across the soluble rock unit; the main criteria referred to by the hydrogeological definition.

The above question reveals a weakness in the standard flank-margin speleogenetic model, which is based on an assumption that the rock sequence is homogenous (Fig. 2A). The references cited in the previous section show that this assumption is not always valid. Moreover, there is a large body of publications that demonstrate significant inhomogeneities, both layered and discordant, and hence the presence of leaky aquifer systems in coastal regions. A simple conceptual setting is presented in Figure 2B, where an aquifer system is depicted consisting of an upper unconfined aquifer bounded from below by a lower confined aquifer, while the aquifers are separated by an aquitard that allows leakage. An aquitard can be heterogeneous in its lateral extent, allowing more significant leakage in certain areas where the vertical conductivity is enhanced due to the presence of fractures or other discontinuities. The obvious result of this circulation pattern would be the formation of truly hypogene caves driven by the leakage of freshwater from the lower aquifer. The aggressiveness would be produced due to mixing of the leaking freshwater with the marine water at the base of the unconfined aquifer, and natural convection effects would be very pronounced due to spatially fixed, steady and efficient supply of freshwater from below.

It is therefore suggested that the flank-margin model should be expanded to account for a multiple aquifer settings. From the perspective of the hydrogeological approach, both epigene speleogenesis and hypogene speleogenesis may develop in coastal carbonates.

CONCLUDING REMARKS

Defined in the most general way, hypogene speleogenesis is the origin of caves in which the cave-forming agency comes from depth, in contrast to epigene speleogenesis in which the cave-forming agency (meteoric recharge and its inherent or soil-derived aggressiveness) originates at the surface. A more specific definition should rely on attributes of the cave-forming agency that are most suitable and efficient for discrimination between epigene and hypogene origins of caves.

Relying on the determination of a source of the aggressiveness in classifying hypogene speleogenesis is the legitimate approach but it is not a methodologically sound and practically efficient one.

The hydrogeological approach and the reference to upwelling groundwater circulation in the definition of hypogene speleogenesis provide a theoretically and methodologically sound basis not only for identifying the type of speleogenesis, but also for spatial and temporal prognosis of hypogene speleogenesis.

Hypogene speleogenesis develops where upwelling groundwater circulation and disequilibrium conditions causing dissolution are supported during a sufficiently long time. It is localized predominantly in discharge zones and/or zones of interaction of groundwater circulation systems of different nature, depth and scales, and it is controlled by peculiarities of the hydrogeological structure, geodynamic evolution and geomorphic development of regions.

In basinal settings, the localization of areas of upwelling circulation across soluble rocks, and hence of hypogene speleogenesis, is determined by the influence on hydrodynamics of the basins topography and configuration, tectonic disruptions, internal uplifts, and lithofacial windows, as well as of endogenous (geodynamic) factors. The role of tectonic faults as cross-formational fluid-conducting structures strongly increases in the lower stories of cratonic artesian basins and in massifs of orogenic regions. The development of hypogenic void-conduit systems is commonly multiphase, determined by major phases of the geodynamic history of the regions.

The patterns and morphology of hypogene caves are determined by the structure of initial porosity, pressurized mode and the upwelling vector of groundwater circulation, specific features of the speleogenetic mechanisms in the conditions of the external conservative control over discharge, as well as by peculiarities of the evolution of a given groundwater circulation system. When hypogenic caves are shifted to the shallow subsurface, their morphology may experience considerable modification by dissolution at the water table and by subaerial mechanisms.

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CONTRIBUTED PAPERS

BAHAMIAN CAVES AND BLUE HOLES: EXQUISITELY PRESERVED FOSSIL ASSEMBLAGES AND TAPHONOMIC INFLUENCES

Nancy A. Albury[1] and John E. Mylroie[2]

In The Bahamas, caves and blue holes provide clues to the geologic and climatic history of archipelago but are now emerging as windows into the ecological and cultural past of islands. Cave environments in The Bahamas alternate cyclically between vadose and phreatic conditions with sea-level change, thereby providing unique but ephemeral fossil capture and preservation conditions.

A diverse assemblage of fossil plants and animals from Sawmill Sink, an inland blue hole on Abaco Island in the northern Bahamas, has revealed a prehistoric terrestrial ecosystem with exquisitely preserved fossil assemblages that result from an unusual depositional setting. The entrance is situated in the pine forest and opens into a flooded collapse chamber that intersects horizontal conduits at depths to 54 meters. The deepest passages are filled with sea water up to an anoxic mixing zone at depths of 14 to 9 meters and into the upper surface fresh-water layer. The collapse chamber is partially filled with a large talus pile that coincides with an anoxic halocline and direct sunlight for much of the day.

During glacioeustatic sea-level lowstands in the late Pleistocene, Sawmill Sink was a dry cave, providing roosting sites for bats and owls. Accumulations of bones deposited in depths of 25 to 30 meters were subsequently preserved by sea-level rise in the Holocene. The owl roost deposits are dominated by birds but also include numerous small vertebrate species that were actively transported by owls to the roost sites.

As sea levels rose in the Holocene, Sawmill Sink became a traditional passive pitfall trap. Significant quantities of surface derived organic material collected on the upper regions of the talus at the halocline where decaying plant material produced a dense layer of peat within an anoxic mixing zone enriched with hydrogen sulfide. Vertebrate species that drowned were entombed in the peat, where conditions inhibited large scavengers, microbial decomposition, and mechanical disarticulation, contributing to the superb preservation of the fossil assemblage in the upper regions of the talus.

Exquisitely preserved vertebrates include the Cuban crocodile Crocodylus rhombifer and the tortoise Chelonoidis alburyorum, flightless birds, and other fossils with ages ranging from 4000 to 1000 years. Because many species are now extinct or no longer occur in The Bahamas, current studies are attempting to resolve the degree to which prehistoric human settlement affected the Bahamian terrestrial flora and fauna.

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HYPOGENE CAVE PATTERNS IN IRON ORE CAVES: CONVERGENCE OF FORMS OR PROCESSES?

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Speleogenesis in iron ore caves may involve generation of porosity at depth with a later surficial phase associated with slope hydrological processes. The earlier phreatic phase results in morphological features similar to but much more irregular at wall and ceiling scale than what is observed in hypogene caves. Processes responsible for the generation of caves do not seem to follow normal karst geochemical paths, but instead occur through bacterially mediated redox reactions.

INTRODUCTION

Caves and small voids in iron-rich rocks have been reported in the Brazilian geological literature since the 19th Century, but only recently, due to the expansion of iron mines, have they been subject to detailed studies. Initial research was performed by American geologists from the United States Geological Survey in the Iron Quadrangle region of southeastern Brazil, with George C. Simmons providing pioneering insights on cave genesis and mineralogy (Simmons, 1963; 1964). Since 2005, with the increase in iron ore prices and the regulatory mandate to assess the significance of any void over 5 m in length, intensive research has resulted in the identification of approximately 3,000 caves, the majority of them being located in the two major iron ore provinces, Carajás ridge in northern Brazil (Amazonia) and the Iron Quadrangle area (Fig. 1). Cave mapping and geospeleological studies have provided new insights on the morphology and genesis of these enigmatic and little known caves.

Figure 1. Location of iron ore outcrops and caves. Besides the better known areas of Carajás and the Iron Quadrangle, iron ore caves are known from the western area of Corumbá, and the south central Espinhaço ridge (Caetité and Conceição do Mato Dentro areas).

GEOLOGY OF IRON ORE CAVES

Iron ore is a generic umbrella term that denotes iron-rich rocks with economic value. These rocks include, besides the original BIF (Banded Iron Formation), a series of heterogeneous alteration products with varying iron content. Although BIF represents the original rock, due to its long term tectonic and weathering history, unaltered BIF is seldom found at or close to the surface. Chemical alteration of BIF is neither a continuous nor a synchronous process, resulting in a complex array of highly heterogeneous rocks with distinct levels of alteration, including a surficial iron-rich breccia cemented by ferruginous matrix known as canga (Fig 2). Alteration of BIF can occur at great depths, as frequently shown by the occurrence of friable high-grade ore deep in open pit mines. Iron compounds are by far the more resistant constituent of BIF, silica together with other elements (carbonates, etc) being more easily removed, resulting in a more porous and friable rock horizons referred to as "pale zones" (McFarlane and Twidale, 1987).

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In the Iron Quadrangle, iron-rich rocks comprise BIF and alteration products of the 2.5 Ga Cauê Formation of the Minas Supergroup (Babinsky et al., 1995). These rocks outcrop at the top of ridges, forming a narrow strip of iron topped by canga, corresponding to the highest elevations in this mountainous area. In the Carajás area, the iron ore belongs to the 2.7 Ga Carajás Formation, Grão Pará Group (Trendall et al., 1998). The local geomorphology comprises a series of irregularly linked flat-topped plateaus capped by canga. As a general rule, the very low rates of denudation in iron outcrops (Spier et al., 2006; Shuster et al., 2012) result in the iron formations occupying the highest elevation terrains.

The majority of caves develop in the irregular contact between altered BIF and canga. Caves entirely in altered BIF or entirely in canga also tend to be common. Caves within BIF, on the other hand, are relatively rare. It must be stressed that the higher frequency of caves in the canga/altered BIF contact may reflect a sampling bias towards erosional exposure of more superficial caves. There is some indication of a significant number of entrance-less voids existing at depth. Table 1 provides brief general data on cave features relative to the lithology.

REMARKS ON THE GENESIS OF IRON ORE CAVES AND THE RESULTING MORPHOLOGY

The existence of substantial voids at great depths in iron areas is well established by numerous proprietary borehole logs from iron exploration studies. Porosity generation and thus speleogenesis may start at an unknown depth, on the order of several hundred meters below the present surface. As already discussed, in both Carajás and the Iron Quadrangle, silica undergoes chemical weathering, yielding a more porous rock and a higher-grade ore. In the Iron Quadrangle some BIFs also contain carbonates, which are also prone to be dissolved away. Irregular, non-connected voids would be created at this early stage.

The chemical processes related to silica and carbonate dissolution are well established (Ford and Williams, 2007; Wray, 2013). The mobilization of iron, however, is a more complex process that has recently been shown to involve bioreduction of Fe (III) by iron reducing bacteria that convert insoluble solid Fe(III) into aqueous Fe(II), allowing for the mobilization of iron and generation of voids (Parker et al., 2013a, b).

As denudation progresses, isostatic rebound will slowly move the caves above the water table towards the land surface. Iron ore caves are located in high elevation areas are relict dry features, hydrologically active in only a limited way; once they get close to the surface, the presence of the very resistant canga cover will further protect the caves from weathering and unroofing. However, these caves will eventually become integrated to the near-surface hydrological processes that occur at the slopes, particularly subsurface flow at the canga/iron ore interface. The original hypogene morphology will tend to be partially obliterated by the connection between chambers, masking or sometimes even completely overprinting the original hypogene features. A schematic model of cave evolution is shown in Figure 3.

Slope hydrological processes will link isolated chambers and will result in a more linear pattern parallel to the slope. In this analysis we will focus in what is interpreted as early hypogene-like features of these caves.

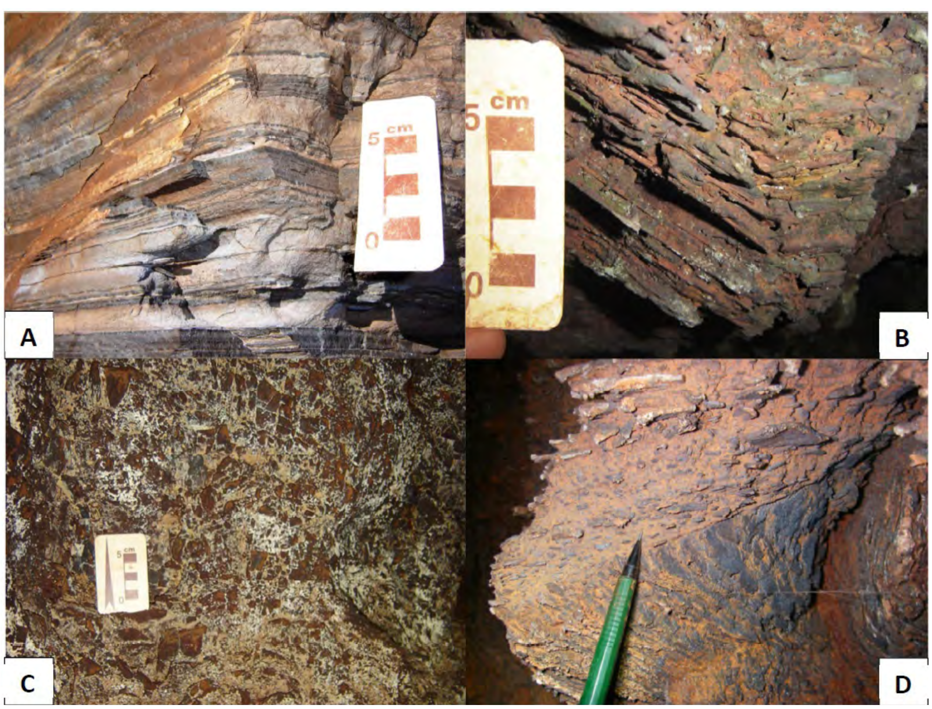

Figure 2. Examples of iron ore rock types. A. Unaltered Banded Iron Formation (BIF) showing silica and iron bands; B. Altered BIF. Silica has been leached. C. Canga showing whitish bacterial colonies. D. Sharp contact between canga (top) and altered BIF (bottom).

Table 1. General morphological features relative to rock type.

Rock Type	Cave Frequency	Morphology	
		Macro	*Micro*
BIF	Rare	Rounded chambers	Polished walls, less irregular at ceiling and wall levels
Altered BIF	Common	Rounded chambers tend to be evident	Irregular at ceiling and wall levels, sharp rock projections are common.
Canga	Common	highly irregular, rounded chambers	Presence of pendants and pillars
Contact Canga/Altered BIF	Very Common	highly irregular, rounded chambers	Display features common to both rock types

Differentiating Between Hypogene and Later Modification Features

In the "hypogene" category we include morphological features that are interpreted as being generated deep below the water table. As previously mentioned, later slope processes result in a more elongated pattern that considerably masks the original morphology. Among these later vadose processes one can include the alignment of the cave with the slope gradient, its close proximity to the surface and the existence of smaller passages linking rounded irregular chambers (Fig. 4). Furthermore, small channels mostly at the contact between floor and walls are quite abundant, but probably represent later generated inlets related to the expansion of the cave along the slope allowing the input of fine grained altered iron sediment. Ubiquitous breakdown is attributed to the very surficial nature of the caves, in which unloading (release) joints favor ceiling collapse.

Hypogene Morphology at Plan View

The morphology at plan view indicates many features that resemble hypogene morphology:

- Absence of an entrance. Iron ore caves usually display an entrance that is much smaller than the remaining inner passages. These entrances are associated with the evolution of scarps, fortuitously intersecting once isolated chambers (Fig. 5). Vertical entrances may provide the sole access to the cave, ceiling breakdown being favored due to the shallow nature of the cave.
- Presence of isolated irregular chambers. These chambers usually present a flat floor and are linked to similar chambers via inclined narrower passages. The linkage between chambers is responsible for generation of some of the longer caves (Fig. 6).
- Cave pattern shows no clear relation to groundwater flow. Pattern may be described as enlarged, isolated vugs.

Furthermore, iron ore caves are devoid of allochthonous sedimentation and display no fluvial sediment assemblages; the clastic sediment sequences are entirely composed of material generated within the cave, being highly homogeneous in terms of composition.

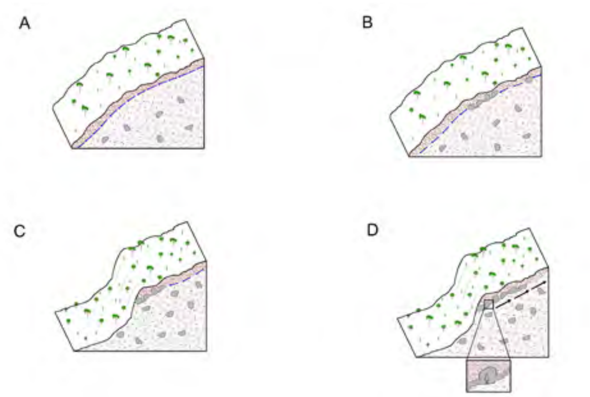

Figure 3. Schematic evolution of iron ore caves. A. Isolated vugs at depth. Denudation and isostatic rebound positions the caves above the water table and closer to the surface. The contact between canga (darker red) and altered ore is a favorable zone for subsurface flow parallel to the slope (blue arrows). B. Caves reach the contact and become integrated to the slope hydrological system. C. Connection between once isolated vugs. D. Scarp retreat intersects a former entranceless cave. Connection between chambers and headward expansion of caves along the contact. At these later stages, the original "hypogene" morphology may become obliterated.

Morphology at Micro Scale

At micro scale the total absence of flow marks is perhaps the most conspicuous feature. No scallops, ceiling or wall dissolution pockets or any other features related to water flow are present at these caves. Pendants and pillars are common, especially when the ceiling is developed in canga (Fig. 7). At a microscale, walls and ceiling tend to be extremely irregular, partially due to the heterogeneous nature of the bedrock (be it canga or altered ore), lacking the polished dissolutional walls typical of carbonate caves.

Figure 4. Profile of a typical iron ore cave (Triangulo cave) elongated parallel to the slope. Map from Pereira (2012).

Figure 5. Example of a once entranceless cave in the Iron Quadrangle. The later connection is suggested by the constricted nature of entrance.

Figure 6. Plan view of a large cave in the Iron Quadrangle showing smaller passages (marked with an "X") connecting once isolated chambers (marked with a circle).

DISCUSSION

Void generation at depth in iron ore areas involves chemical/microbiological removal and transformation of iron constituents (besides silica), resulting in a highly irregular array of non-connected voids of various sizes, commonly intersected by boreholes at depths up to several hundred meters below the surface. Speleogenetic processes present similarities to hypogene settings in the sense that initial rock porosity is generated at depth, with no influence from surface processes. Sluggish water flow and lack of connectivity between pores result in irregular isolated chambers, similar to flank margin cave morphology (Waterstrat et al., 2009). Even taking into account the very long timescales (on the order of hundreds of million years) available for speleogenesis in iron ore regions, the more chemically resistant nature of iron formations result in much smaller voids and caves when compared to carbonate settings.

Iron ore caves do not display the integrated nature of many hypogene caves, with ramiform or spongework caves (sensu Palmer, 1991) being unknown, although the small size of the caves would make the recognition of such pattern difficult. Unlike many typical hypogene caves (Klimchouk, 2007) iron ore caves are not evidently linked to rising ground water paths, and thus there is no clear relationship between caves and ground water routes.

It appears that the final morphology of iron ore caves is heavily affected by surficial processes. Unlike hypogene caves, these caves seem to become integrated with slope hydrological processes, resulting in an amalgamation of once isolated chambers, the final morphology displaying a much more linear pattern that parallels the slope.

CONCLUSIONS

The first stage of speleogenesis in iron ore caves bears similarities to hypogene processes, in that it develops at depth, away from surficial processes. The morphology at plan scale resembles slow-flow non-integrated hypogene caves such as flank margin caves. Sluggish water flow environment, dominance of chemical and microbiological processes and decoupling in relation to water flow routes are common features, although iron ore caves evolve through much longer timescales. Unlike hypogene settings, in which an active hypogene cave will eventually become detached from groundwater systems, the later evolution of iron ore caves result in a drastic transition between a sluggish phreatic deep environments towards a shallow vadose environment subject to hydrological slope processes.

Figure 7. A very irregular chamber at an iron ore cave in the Iron Quadrangle. The ceiling, developed in canga, displays numerous pendants and pillars. Photo by Ataliba Coelho.

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EARTH TIDE, A POTENTIAL DRIVER FOR HYPOGENIC FLUID FLOW: OBSERVATIONS FROM A SUBMARINE CAVE IN SW TURKEY

C. Serdar Bayari[1] and N. Nur Ozyurt[1]

Initiation and development of karstification requires a continuous flushing of pore water in equilibrium with carbonate minerals. Under confined flow conditions, the energy required for pore water transport is supplied by external pressure sources in addition to the by earth's gravity. Earth tides and water loads over the confined flow system are the main sources of external pressure that drives the pore water. Earth tides, created by the sum of the horizontal components of tide generation forces of moon and sun, causes expansion and contraction of the crust in horizontal direction. Water load on top of the confined flow system causes vertical loading/unloading and may be in the form of recharge load or ocean loading in the inland and sub-oceanic settings, respectively. Increasing and decreasing tide generating force results in pore water transport in the confined system by means of contraction and expansion, respectively. Since these forces operate in perpendicular directions, pore water flushing by earth tides becomes less effective when water load on top of the confined flow system increases. Temporal variation of freshwater content in a submarine cave is presented as an example of groundwater discharge driven by earth tides and recharge load.

INTRODUCTION

Increasing number of evidences reveal that hypogenic fluid flow has a more important role than previously thought in initiation of karstification and development of karst aquifers. However, importance of hypogenic karst development, particularly in deep-burial or mesogenetic environments is also debated because the pore water which becomes saturated with respect to carbonate minerals must be renewed in order for sustaining further dissolution (e.g. Klimchouk, 2000 and the references therein). Therefore, one of the critical issues regarding the hypogenic karst development is the source of energy required to drive the pore water. Earth's gravity and external pressure are the major sources of energy by which the groundwater water is transported. Earth's gravity determines the hydrostatic pressure (Ps) both in confined and unconfined porous media whereas, the pressure exerted on the medium from outside provides an additional energy (excess pressure, P*ex*) for pore water transport in the confined systems. Flow systems bounded by lithological units with remarkably lower hydraulic conductivity constitute the confined flow systems. Moreover, many unconfined systems include vertical hydraulic conductivity contrasts that make the system confined at varying rates. Consequently, confined flow appears to be more common than the completely unconfined flow systems at global scale. In the following, we present the fundamental equation of flow in confined porous medium and the common sources of P*ex*. Then, we explain how the external pressure sources like earth tides and recharge load over the aquifer affects the P*ex* in confined flow systems. Finally, we show briefly how the interplay of these P*ex* sources can affect the temporal aquifer deformation which in turn results in temporal fluctuations in the discharge of aquifer. Details of the case study will be presented elsewhere.

DRIVERS OF FLOW IN CONFINED SYSTEMS

An extended version of Darcy's equation (e.g. Domenico and Schwartz, 1997) can be used to describe the combined effect of hydrostatic pressure (P*s*) and excess pressure (P*ex*) gradients on mass flux in partly or fully confined flow systems (Fig. 1),

$$q = K\frac{\partial}{\partial h}\left(z + \frac{Ps}{\rho g} + \frac{Pex}{\rho g}\right) \tag{1}$$

where, q is groundwater flux, K is hydraulic conductivity, h is head, z is head over a datum, ρ is density of water, and g is the earth's gravitational acceleration constant. As is inferred from Equation 1, any load exerted on a fully or partially confined aquifer would result in movement of groundwater because of the P*ex* created. Fluctuation of groundwater level in artesian wells due to a nearby railroad activity is a classic example of this effect on potentiometric level. Seismic shocks, storms, floods, recharge over the surface of aquifer, ocean loading on top of the aquifer, atmospheric pressure and earth tides are other examples of external loads exerted on confined aquifers (e.g. Domenico and Schwartz, 1997). These loads are hard to sense since they are applied slowly. However, they cause measureable changes on P*ex*. For example, recharge load (RL) due to seasonal precipitation input over Amazon Basin leads to ca. 8 cm of vertical fluctuation at continental scale (Tapley et al., 2004; Bevis et al., 2005). Vertical crustal deformation due to RL is slow because of the slow loading (recharge) and unloading (drainage) processes that are realized usually at seasonal time scale.

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Earth tides are another source of Pex in confined systems. All celestial bodies exert a natural force of attraction to other masses, known as gravity. The gravity between any two bodies is directly proportional to the product of their masses and is inversely proportional to the square of the distance between them. The gravity exerted by moon and sun over the earth is the primary reason of the earth and ocean tides. The effects of other celestial bodies on earth and sea tides are negligible. While both of the earth and ocean tides are the result of gravitational forcing of the moon and sun, they have different properties. Sea tides are affected also by other factors like currents and bathymetry and, their range of fluctuation may reach about 16 m (O'reilly et al., 2005). Sea tides causes movement of huge water masses which, in turn, may result in a substantial ocean loading over the underlying geologic material. On the other hand, earth tide causes the elastic deformation of the lithosphere at periods of about 12 hours and longer. Elastic deformation caused by earth tides is in the form of bulges and depressions with a magnitude of less than a meter (mostly around 0.3 m - 0.5 m).

The stress imposed by earth's tides on a confined (e.g. Ferris et al., 1962; Gieske and De Vries, 1985) and unconfined flow systems (e.g. Rojstaczer and Riley, 1990) cause micron-level deformation (Maucha and Sarvary, 1970) that lead to head oscillations (Hobbs and Fourie, 2000), discharge rate fluctuations (Maucha and Sarvary, 1970; Williams, 1977), radon flux fluctuations from crust in karstic (Barnet et al., 1997) and hard rock aquifers (Maréchal et al., 2002). Moreover, periodic oscillations in submarine natural gas discharges (Luyendyk et al., 2003), geyser activities (e.g. Rinehart, 1972), and oceanic vent fluids (e.g. Schultz et al., 1996) are likely to be linked with earth-tidal elastic deformation.

Simply stated, tide generating force (TGF) causes horizontal expansion (high tide) and compression (low tide) of the crust at about 12 hours and longer time scales (Figure 1). On the other hand, RL causes vertical compression (high load) and expansion (low load) at seasonal scale. Since these forces are perpendicular to each other, the ratio of these vectors determines the magnitude of P*ex*. To ease the interpretation of combined effect of TGF and RL on the P*ex*, a parameter called "aquifer deformation factor (ADF) is described as the ratio of TGF to RL (Fig. 1). Since both forces are in N units, ADF is a dimensionless quantity. Increasing and decreasing TGF and RL causes increasing aquifer contraction and expansion, respectively.

TEMPORAL VARIATION OF EXCESS PRESSURE

Tide Generating Force

Anywhere on the surface of earth, the tidal force exerted by the moon's (or sun's) gravitational attraction can be separated into two force components, one perpendicular to and the other tangent to the earth's surface. This second component is the main driver of earth tides. Earth tides created by the tide generating force (TGF) of the horizontal gradient component of the joint gravity field of the moon and sun that account for the 1/3 and 2/3 of the combined force of attraction on earth, respectively. The horizontal component of attraction force is zero at the points on the earth's surface directly beneath the moon (time of full moon) and on the opposite side of the earth from the moon (time of new moon).

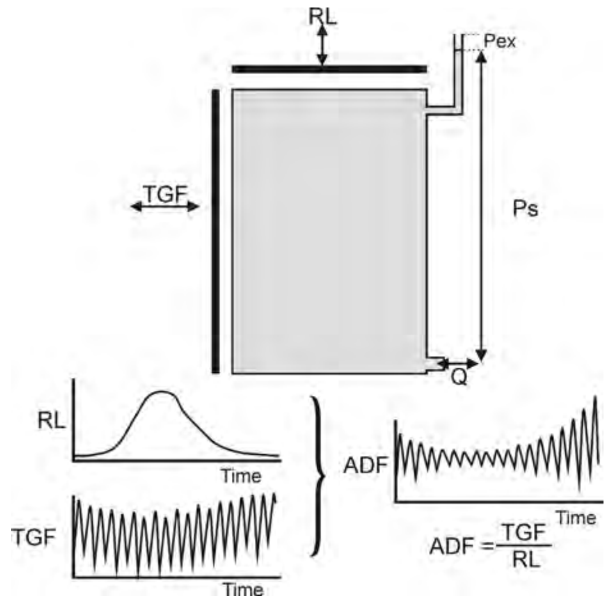


Figure 1. Vertical (RL) and horizontal (TGF) external forces affecting a confined aquifer.

The horizontal component of the TGF exerted by sun or moon is calculated by the following equation (Butikov, 2002),

$$TGF_i = \left\{ -\frac{3}{2} \frac{GmM_i r_E}{d_i^3} \right\} \left\{ Sin\left(2\pi \frac{t}{\tau_i} + p_i \right) \right\} \qquad (2)$$

where TGF_i is horizontal component of tide generating force (N) exerted on 1 kg mass (m) on the earth's surface by the celestial body i (= sun or moon), M_i is the mass (kg) of i, G is the universal gravitational constant, d is the distance (m) to sun or moon, r_E is the radius of earth (m), *t* is the day number for which TGF is calculated, ι_i is the orbital period of celestial body *i* (days), p_i is the phase shift (number of days passed between latest maximum phase and the first day of calculations). The earth-moon and earth-sun distances for the geocentric coordinates of the area of interest for any given time can be obtained by using an ephemeris (i.e. celestial coordinates) software (e.g. Alcyone Ephemeris, 2013). Figure 2 shows an example of temporal variation of total TGF calculated for 36.25 °N, 29.50 °E coordinates during a period starting from November 1st, 2004. The TGF exerted by sun shows a seasonal variation whereas the TGF exerted by moon has bi-weekly periods. As seen in Figure 2, total TGF is

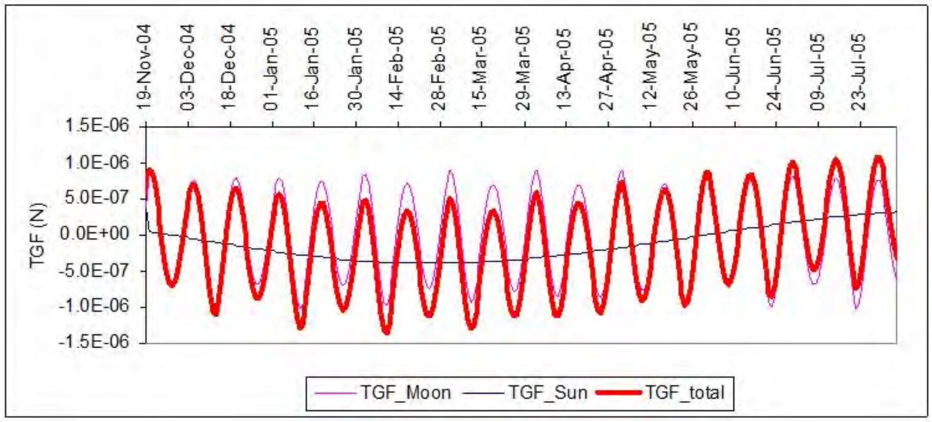


Figure 2. Temporal variations of total TGF, moon's TGF and sun's TGF.

the sum of both TGFs. The more positive the values of total TGF vector, the more the crust contracts (high tide). Similarly, the more negative the values of total TGF vector, the more the crust expands (low tide).

Recharge Load

Another source of external force that may be exerted on a confined aquifer on land is the recharge load that occurs mainly due to precipitation falling over the aquifer. Large water mass movements that occur due to sea tides may also exert substantial loads (i.e. ocean loading) over the submarine aquifers. Magnitude of the RL on land depends on the magnitude of recharge (e.g. precipitation, flooding) and recharge holding capacity of the ground overlying the confined unit. Systems with high recharge capacity and low surface/subsurface drainage capability possess high RL. The timing of the RL depends usually on the regional climate. Since the humidity in the polar and desert areas is too low, most of the precipitation falls in temperate and tropical zones where seasonality of recharge depends on the latitude. Because of the obliquity of earth's rotation axis, solar radiation arriving at unit surface area changes seasonally in the temperate zones whereas such an annual change over the tropical zone is much more limited. Eventually, the wet season of the temperate zones of northern and southern hemispheres realizes around northern and southern winters, respectively. Tropical climate is characterized by two wet periods which roughly occur around June and October. Figure 3 exhibits the seasonal variation of conceptual recharge loads (RLs) under temperate and tropical climates between 19 November 2004 and 18 November 2005. The RLs shown in the figure were normalized subjectively to range between 0.5 N (ca. 0.05 kg) and 2 N (ca. 0.20 kg) to prevent division error in the following ADF calculations. Temporal TGF variation shown in the figure may be regarded as representative for 36°N (i.e. representative of northern temperate zone), 36°S (i.e. representative of southern temperate zone) and 12°(i.e. representative of tropical zone) latitudes and at 29°E longitude because the difference among the TGFs calculated for different latitudes is less than +/- 0.17 %. At any given time, there is a maximum 14.5 days of phase shift between the TGFs calculated for a longitude and its antipod. Therefore, temporal TGF variation shown in Figure 3 may be regarded as a global TGF which is valid anywhere with an error of +/- 14.5 days.

Aquifer Deformation Factor

The ADF, as defined above, is the ratio of TGF to RL. Figure 4 shows temporal variation of the ADFs for TGF at 36°N divided by RLs at temperate (36°N, 36°S) and tropical (12°N) climates. All ADFs follow a similar temporal trend during which their am-

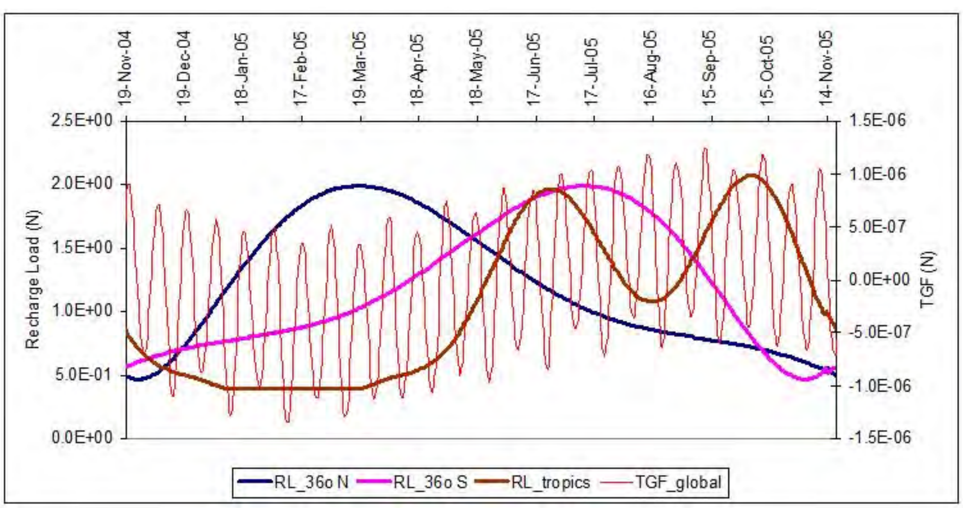


Figure 3. Seasonal variation of the conceptual recharge loads under temperate and tropical climates.

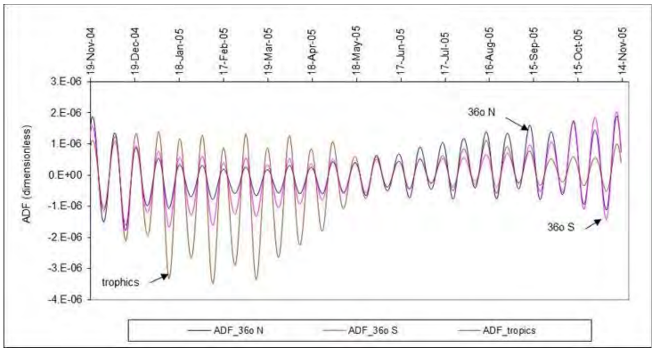


Figure 4. Temporal variation of conceptual ADFs for different recharge loads.

plitude change depending mainly on the respective RLs. It is apparent that the magnitude of ADF is controlled globally both by the TGF and local RL. However, while the portrait shown in Figure 4 seems simple for the inland aquifers, the earth-tidal loading process may be much more complicated in submarine confined aquifers because the ocean loading caused by sea tides varies spatio-temporally. Besides, the absolute magnitude of ADF depends on the absolute magnitudes of RF and TGS. The absolute magnitude of RL can be estimated from water budget calculations that give the temporal "water load" on top of the confined aquifer. However, estimation of the absolute TGF for the aquifer of interest is a more difficult task because the absolute mass of the aquifer is difficult to know.

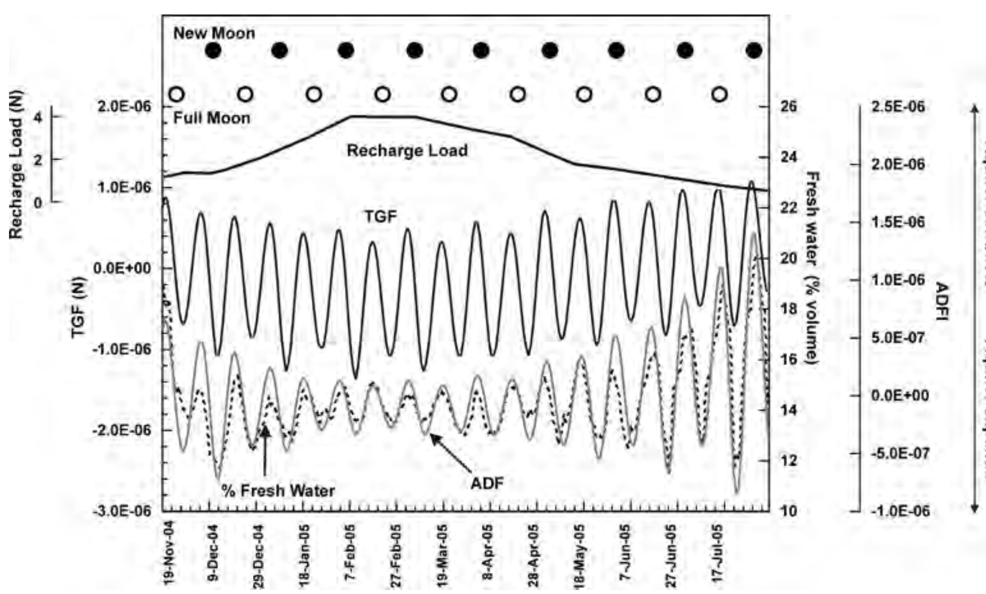
As seen on Figure 4, in all climates, largest ADF oscillations are observed when recharge load is small (i.e. in dry period). The aquifer is expansion-dominated around January when the sun is closer to earth. Largest aquifer compression is observed in the tropical zone between November and May when the crust is expanded more and the recharge load is at minimum.

PUMPING OF GROUNDWATER BY EARTH TIDES AND RECHARGE LOAD

Long-term salinity observations carried out in a submarine cave in SW Turkey provide a nice example of how ADF, governed by earth tides and the recharge load cause temporal variation of groundwater discharge into the sea. The observations were obtained by a data logger located at the ceiling at -20 m below sea level of the cave. The cave is developed in Mesozoic carbonates along the seawater / freshwater interface that extends from – 4 m to – 80 m below sea level. Mediterranean type climate with warm-wet winters and hot-dry summers dominate the area where the mean annual precipitation is around 0.9 m. The recharge area extends from sea level to more than 3000 m above sea level and comprises almost entirely of carbonate rocks with extensive karst development (Elhatip and Günay, 1998). Specific conductance and temperature observations in the cave have been conducted to understand temporal dynamics of submarine groundwater discharge. These signals were used to determine the specific conductance of sea water in the cave which in turn was converted to freshwater content of seawater (Figure 5).

The freshwater content supplied by the karst aquifer shows a sinusoidal pattern with amplitude that varies temporally. The magnitude of freshwater content ranged from 11 % (by volume) to 21 % during the observation period. Interestingly, the mean freshwater content in the midst of wet season (i.e. December-May) is lower than mean freshwater content during the dry season that extends from June to November. Since this part of the Mediterranean Sea is far from Atlantic Ocean, sea tides cause limited sea level fluctuations (+/- 0.1 m; e.g. Aviso, 2013). Therefore, ocean loading is not an accountable external pressure source. Barometric pumping, resulting from the atmospheric pressure gradient between the western and eastern ends of Mediterranean Sea does not show temporal cycles so that observed freshwater content variations in the cave cannot be attributed to this phenomenon. The cycles of increasing and decreasing freshwater content is in

Figure 5. Comparison of the temporal variations of percent fresh water discharge from a submarine cave with those of TGF, RL and ADF.



phase with the cycles of TGF calculated for this area. However, the amplitude of cycles of freshwater content and those of TGF are not identical although, their long-term trends are similar. On the other hand, temporal signal of freshwater content exhibits an almost perfect match to the ADF calculated by TGF to RL ratio. The TGF operates in horizontal direction and increasing and decreasing values of TGF implies aquifer contraction and expansion, respectively. The magnitude of elastic deformation applied on aquifer by TGF is reduced by the RL which operates in vertical direction. As a consequence, the discharge of karst aquifer driven by TGF is reduced during the wet season when the elevated RL limits the aquifer deformation.

CONCLUSIONS AND OUTLOOK

Results of this study reveal that i) even a seemingly-unconfined karst aquifer can behave confined because of the extreme gradients of hydraulic conductivity, and ii) flow of groundwater in karst aquifers can be driven by earth tides. Since the recharge load reduces the aquifer's response to tide generating force, groundwater flow driven by earth tide appears to be more effective in low-recharge areas or during the dry period. Future studies should be focused on similar studies at different geographic/climatic regions of the world to validate and enhance the results of this study.

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PRELIMINARY CONSIDERATIONS ON HYPOGENE MORPHOLOGY IN TOCA DA BOA VISTA E TOCA DA BARRIGUDA CAVES, NORTHEASTERN BRAZIL

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The Toca da Boa Vista and Barriguda caves are located in Northeastern Brazil. They occur in the Neoproterozoic carbonates (limestones and dolomites) of the Salitre Formation, located at Irecê Basin. This set of rocks occurs within the São Francisco Craton, a region that was not affected by the Brasiliano-Pan-African orogeny (Pedreira et al., 1987). The caves occur at a distance of approximately 300 m apart and there is a possibility of a link between them, but so far this has not been proven. Toca da Boa Vista has about 108 km of mapped passages and is therefore the largest cave in South America. Toca da Barriguda is smaller and has about 32 km of mapped galleries.

The architecture of the Toca da Boa Vista and Barriguda caves present both a 2D network and spongework type (Auler, 2009). The control of the conduits is related to faults, fractures and axial planes of antiforms. The general configuration of the caves seems to follow the Pacuí riverbed that has its channel located about 1km southeast. The origin of these hypogenic caves was first postulated by Auler & Smart (2004), who described some hypogenic features and reported a acid source (H_2S) coming from existing pyrite in carbonates to explain the corrosion and dissolution of carbonate rocks. Klimchouk (2009) wrote about the need to investigate deeper this issue. He drew attention to the apparent feeders presence coming from the lower aquifer as well as to the importance of determination of the source of acidity, since the amount of pyrite present doesn't seem to be significant for the origin and development of the caves by hypogenic speleogenesis.

Although the origin and development of the caves are still under discussion, abundant hypogenic forms are present. Feeders, rising wall channels, half ceiling tubes, half wall tubes, ceiling cupolas, convection cupolas and wall niches are the major forms found. The linear geometry of caves suggests that they have a structural control. In addition, cavities generated at Toca da Boa Vista and Barriguda caves seem to follow the same stratigraphic level, as well as existing permeable structures such as fractures, faults and axial planes of antiforms. The process of ascending flow through these structures has resulted in the opening of the cavities by hypogenic dissolution as well as the collapse of blocks caused by the lack of sustainability of the layers generated by the voids left by the dissolution. Outlets that would flow to levels above were not found. The origin and evolution of the cave system, however, needs further investigation.

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FACIES ANALYSES AND DIAGENETIC EVOLUTION OF THE HYPOGENE TOCA DA BOA VISTA/TOCA DA BARRIGUDA CAVE COMPLEX, SALITRE FORMATION (IRECÊ BASIN, BRAZIL) – PRELIMINARY RESULTS

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The present study investigates the lithofacies and processes such as karstification in the karst province of the Neoproterozoic Una and Bambuí groups, located in the central-eastern part of Brazil. This province comprises several cave systems in carbonate rocks (Karmann and Sanchéz, 1979), which include the Toca da Boa Vista and Barriguda caves, considered the largest hypogenic caves in South America (Auler and Smart, 2003). These caves were formed mainly in dolomites of the Salitre Formation, which was deposited in a shallow marine environment in an epicontinental sea (Medeiros and Pereira, 1994). The Salitre Formation in the cave area comprises laminated mud/wakestones, intraclatic grainstones, oncolitic grainstones, oolitic grainstones, microbial laminites, colunar stromatolites, trombolites and fine siliciclastic rocks (marls, shales and siltites). A thin layer and nodules of chert also occur, which sometimes are seen to fill faults and fractures. Phosphate deposits are also found. Our study focuses on cave mapping, thin-section description, micro-tomography, isotopic analysis, and determination of petrophysical properties of different lithofacies. Our preliminary data indicate several complex events of porosity increase and destruction, which led to the formation of the present karts system.

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MODELING SPELEOGENESIS USING COMPUTATIONAL FLUID DYNAMICS: POTENTIAL APPLICATIONS TO HYPOGENE CAVES

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Numerical models of speleogenesis typically simulate flow and dissolution within single fractures or networks of fractures. Such models employ fracture flow and pipe flow equations to determine flow rates and only consider average velocities within each fracture segment. Such approximations make large scale simulations of speleogenesis tractable. However, they do not allow simulation of the formation and evolution of micro- or meso-scale cave passage morphologies. Such morphologies are frequently studied within a field setting and utilized for the interpretation of the speleogenetic processes that formed the cave. One classic example is the formation of scallops in cave streams with turbulent flow. Scallops are used to interpret past flow velocities and directions. However, a recent analysis of the theory of limestone dissolution in turbulent flow conditions suggests a discrepancy between theory and reality concerning the formation of limestone scallops (Covington, in review). Similarly, the only attempt to numerically simulate flute formation in limestone found that the flute forms were not stable (Hammer et al., 2011). Motivated by these puzzles, we are developing a computational fluid dynamics (CFD) framework for the simulation of the evolution of dissolution morphologies.

While this project was initially conceived to better understand dissolution in turbulent flow, the tools being developed are particularly well-suited to examine a variety of other questions related to cave morphology on the micro- and meso-scales. There has been significant recent discussion about the interpretation of features that are diagnostic of hypogenic or transverse speleogenesis, such as the morphological suite of rising flow defined by Klimchouk (2007). Other authors have suggested that such forms can be found in a variety of settings where confined flow is not present (Mylroie and Mylroie, 2009; Palmer, 2011). We propose that simulation of such forms using a CFD speleogenesis code will allow a more complete understanding of the connections between process and form, because in such simulations the processes occurring are well-known, well-defined, and also can be adjusted within controlled numerical experiments, where relevant parameters and boundary conditions are systematically varied.

The CFD framework we are developing is based on the Lattice Boltzman method (Chen and Doolen, 1998), which is a popular technique for modeling the mechanics of complex fluids, including fluid mixtures, reactive transport, porous media flow, and complex and evolving domain geometries. With this framework it is straightforward to simulate many of the processes occurring in hypogene settings, including complex fluid flows, dissolution, solute and heat transport, and buoyancy-driven flow. Furthermore, this modeling framework allows these processes to be coupled so that their interactions and feedbacks can be explored. With the suite of capabilities provided by this framework, we can begin to numerically simulate the processes occurring in hypogene speleogenesis, including the driving mechanisms and the role of buoyancy-driven flow and its relationship with the morphological suite of rising flow. In the spirit of a workshop, this work is presented as in-progress, in the hopes that it will stimulate discussion on potential applications of the model being developed.

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A REVIEW ON HYPOGENE CAVES IN ITALY

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Although hypogene cave systems have been described since the beginning of the 20th century, the importance in speleogenesis of ascending fluids that acquired their aggressiveness from in-depth sources has been fully realized only in the last decades. Aggressiveness of waters can be related to carbonic and sulfuric acids and the related corrosion-dissolution processes give rise to different types of caves and underground morphologies.

The abundance of hydrothermal springs and associated travertine deposits, and the widespread interaction between volcanic or sub-volcanic phenomena and karst in many sectors of the Italian peninsula are a strong evidence of hypogene speleogenesis. Furthermore, researches on secondary minerals have allowed to discover hypogene caves formed by highly acidic vapors in subaerial environments, also showing that most of these caves have extremely rich mineral associations.

Despite this, until the late 1980s the only known important cave systems of clear hypogene origin in Italy were considered to be the ones hosted in the Frasassi Canyon and Monte Cucco, in which important gypsum deposits undoubtedly showed that sulfuric acid played an important role in the creation of voids (Galdenzi, 1990, 2001; Galdenzi & Maruoka, 2003; Menichetti et al., 2007). Afterwards many other caves were categorized as formed by the sulfuric acid speleogenesis throughout the entire Apennines. Following the broad definition of hypogene caves by Palmer in 1991, and the even more general one of Klimchouk in the last decade (Klimchouk, 2007, 2009), the number of caves considered of hypogene origin in Italy has grown rapidly. Figure 1 shows the hypogene karst systems of Italy, including, besides the well-known and published ones, also the known and less studied, and presumed hypogene cave systems (see also Table 1).

More recently, in some of these caves detailed studies have been carried out including geomorphology, mineralogy, and geochemistry. Sulfuric acid caves are known from many regions along the Apennine chain (Tuscany, Umbria, Marche, Latium, Campania, Calabria) (Forti, 1985; Forti et al., 1989; Galdenzi and Menichetti, 1989, 1995; Galdenzi, 1997, 2001, 2009; Galdenzi et al., 2010; Piccini, 2000; Menichetti, 2009, 2011; Mecchia, 2012; De Waele et al., 2013b), but also from Piedmont, Apulia, Sicily (Vattano et al., 2013) and Sardinia (De Waele et al., 2013a). In this last region ascending fluids have also formed a hypogene cave in quartzite rock. Oxidation of sulfides can locally create hypogene cave morphologies in dominantly epigenic caves, such as in the Venetian forealps (this cave is not shown in Figure 1, being largely epigenic in origin) (Tisato et al., 2012). Ascending fluids have also created large solution voids in Messinian gypsum beds in Piedmont, and these can be defined hypogene caves according to the definition by Klimchouk (Vigna et al., 2010). Some examples of hypogene cave systems due to the rise of CO_2-rich fluids are also known in Liguria and Tuscany (Piccini, 2000). In the Alps and Prealps (Lombardy), some ancient high mountain karst areas exhibit evidences of an early hypogene origin, deeply modified and re-modeled by later epigenic processes. Hypogene morphologies are thus preserved as inactive features, and it is often difficult to distinguish them from epigenic ones.

At almost twenty years distance from the first review paper on hypogene cave systems in Central Italy by S. Galdenzi and M. Menichetti (1995), we give a review of the state-of-the-art knowledge on hypogene caves actually known from the whole of Italy.

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Table 1. Hypogene cave systems of Italy ordered in well-known (studied), known and presumed ones. For location see Figure 1.

Cave system	Studied	Known	Presumed
1 Valle di Scerscen		X	
2 Valmalenco		X	
3 N Grigna		X	
4 Tremezzina		X	
5 Lezzeno			X
6 Mt Fenera			X
7 Moncalvo	X		
8 Grotta della Guerra			X
9 Scogli Neri		X	
10 Monsummano	X		
11 Mt Pisani		X	
12 Tassare-Mt Nerone	X		
13 Mt Rotondo	X		
14 Mt Cucco	X		
15 Frasassi	X		
16 Faggeto Tondo	X		
17 Mt di Campiglia			X
18 Mt. Civitello	X		
19 Mt Cetona		X	
20 Roselle		X	
21 Montecchio	X		
22 Parrano	X		
23 Pozzi della Piana	X		
24 Triponzo	X		
25 Acquasanta Terme	X		
26 Cittareale		X	
27 Buco del Pretaro	X		
28 Mt Soratte		X	
29 Pozzo del Merro	X		
30 Subiaco		X	
31 Fiume Coperto	X		
32 Mt Lattari		X	
33 S. Cesarea Terme	X		
34 Capo Palinuro	X		
35 Mt Sellaro	X		
36 Cassano allo Ionio	X		
37 Terme Luigiane		X	
38 Phaff	X		
39 S. Barbara	X		
40 Corona 'e Sa Craba	X		
41 Monte Inici	X		
42 Acqua Fitusa	X		
43 Mt Kronio	X		

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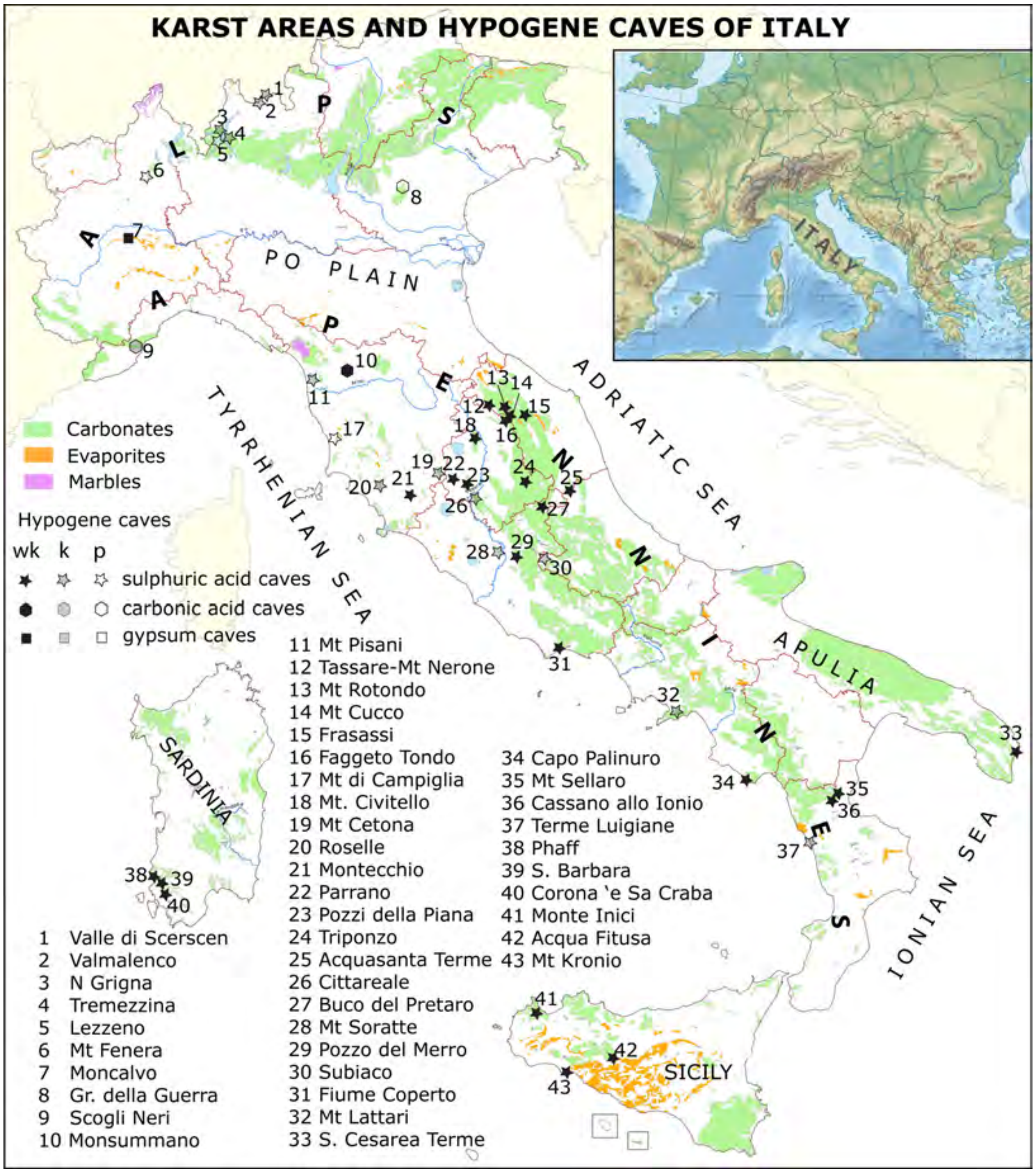

Figure 1. Italian karst areas and the hypogene karst systems (modified from Sivelli and De Waele, 2013, Speleologia 68, special issue printed for the 16th ICS Brno). wk = well known, k = known, p = presumed (GIS elaboration by M.L. Garberi).

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SULFURIC ACID WATER TABLE CAVES (GROTTE DU CHAT / ACQUA FITUSA /BAD DEUTSCH ALTENBURG + KRAUSHÖHLE)

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Sulfuric acid caves can display a variety of forms, from 3D maze systems, to isolated chambers, and more or less maze-like water table caves. Most of the voids are normally generated at or immediately above the water table, where condensation-corrosion processes are dominant, creating a set of characteristic meso- and micromorphologies. This paper deals with the description of four very typical sulfuric acid water table caves: the Grotte du Chat in Provence (France), the Acqua Fitusa Cave in Sicily (Italy), and the Bad Deutsch Altenburg and Kraushöhle caves in Austria.

INTRODUCTION

Until recently, most limestone caves were believed to form by dissolution of rock by meteoric water that acquired its acidity by surface-derived CO_2. These epigenic caves differ from the so-called hypogenic ones by the direction of the water flow, i.e. descending in the first and ascending in the second case (Klimchouk, 2007, 2009) or by the origin of acidity, generated at depth for the latter (Palmer, 1991, 2011; Tisato et al., 2012). In more recent years, an increasing number of caves thought to be epigenic in origin have revealed being caused by rising flows or by waters that acquired their aggressivity at depth (Klimchouk, 2007, 2009). Sulfuric acid caves are the most interesting and best studied among these.

Sulfuric acid caves have been known for a long time in Europe (e.g., Socquet, 1801; Hauer, 1885; Principi, 1931; Martel, 1935), but their detailed genesis was described in American caves much later (Morehouse, 1968). The first sulfuric acid speleogenesis (SAS) model was published by Egemeier (1981) based on observations in Lower Kane Cave in Wyoming. In this cave, H_2S degasses from rising groundwater in the underground void. This gas oxidizes in the cave air or on the moist walls forming sulfuric acid that reacts with the carbonate rock, giving rise to gypsum precipitation. Many sulfuric acid caves are known around the world, such as the famous Carlsbad and Lechuguilla Caves in New Mexico (Hill, 1987, 1990; Polyak et al., 1998; Hose et al., 2000; Engel et al., 2004, Calaforra and De Waele, 2011), la Cueva de Villa Luz in Mexico (Hose and Pisarowicz, 1999), and Frasassi caves in Italy (Galdenzi and Menichetti, 1995; Galdenzi and Maruoka, 2003). Most of these SAS caves are water table caves, i.e. formed along the more or less horizontal plane of the sulfuric groundwater level. Fluctuations of this level can cause these karst systems to exhibit stacked cave levels, mimicking the rise or fall of groundwater.

This paper deals with the detailed description of four of these SAS water table caves, one in France, one in Sicily, and two in Austria (Fig. 1).

Figure 1. Location of the four described caves.

SULFURIC ACID SPELEOGENESIS (SAS)

Sulfuric acid caves are mainly formed above the water table by the abiotic and/or biotic oxidation of H_2S deriving from a deep source (Galdenzi & Maruoka, 2003). H_2S can be produced at depth by volcanic activity, reduction of sulfates (mainly gypsum), or hydrocarbons and is brought to the surface along deep tectonic structures. The origin of the sulfur can usually be ascertained using its stable isotope signature and that of its possible sources (Onac et al., 2011). The oxidation of H_2S produces sulfuric acid that immediately reacts with the limestone host rock producing replacement gypsum and carbon dioxide. This last gas then can dissolve in water again and increase its aggressiveness even more. Also the local oxidation of sulfides such as pyrite,

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often present in limestone sequences, can cause the formation of sulfuric acid, speeding up the dissolution of the carbonate rock (Tisato et al., 2012). The reaction of sulfuric acid with other minerals (e.g. clays) causes the formation of other sulfates such as jarosite, alunite, basaluminite, etc. Some of these minerals, especially those containing potassium, can be dated with the K/Ar and Ar/Ar methods (Polyak et al., 1998). Also gypsum can be dated using the U/Th method (Sanna et al., 2012). The ages of minerogenesis exactly corresponds to the age of cave formation, when the SAS process was active.

Sulfuric acid caves are thus often intimately related to the contact zone between the water level, from which H_2S rises, and the air. Enlargement of the voids mainly happens by condensation-corrosion processes in a highly acidic environment (Audra et al., 2007). These processes are greatly enhanced in the presence of thermal differences between the rising (thermal) waters and the cave walls and atmosphere. Dissolution of carbonate rock in these conditions is extremely fast compared to normal epigenic caves and can cause the formation of sizeable cavities in probably only a few thousands of years.

MORPHOLOGY OF SULFURIC ACID WATER TABLE CAVES

The most typical form of sulfuric acid cave is the water table cave (Audra et al., 2009a, 2009b), with a more or less horizontal development mimicking the former or still active level of the rising more or less thermal and acidic fluids. These H_2S containing waters are discharged through deep fractures that are enlarged in the oxidation (vadose) zone. These feeding fissures (Figs. 2A-C-D, 3A, and 4D) tend to close going downward, because dissolution, and thus enlargement, is most efficient in the altitudinal area in which H_2S oxidizes forming sulfuric acid. This creates sulfuric acid chimneys (Fig. 4B), which can evolve into discharge slots inside well enlarged cavities and finally can become perched above the dropping water table. Most of the cave enlargement occurs above these slots in vadose conditions by condensation-corrosion processes. Upstream of these discharging slots the caves tend to taper out rapidly.

Sulfuric acid water table caves are characterized by a perfectly horizontal cave level developing along the fractures through which acidic fluids were discharged. This can result in a more or less elongated anastomotic cave passage with the discharging slots along its path and right up to the blind end, or in maze caves where fluids rise along a set of fractures. This type of maze is due to rapid re-routing producing new passages that eventually merge by integration (Osborne, 2001). Inclination of these cave passages is usually less than 1%.

Sulfuric acid water table caves are characterized by a typical set of medium- to small-scale morphological features related to the action of sulfuric acid and the related condensation-corrosion phenomena (Galdenzi, 2001; Audra, 2008).

Figure 2. Grotte du Chat: A. Corrosion table and discharge slot (with person inside) along a subvertical fracture; B. Intense corrosion along the walls and boxwork on the ceiling; C. Gypsum crusts still present in convection niches above the corrosion table. Note the legs of the person are in the discharge slot; D. Corrosion table in a wide anastomotic passage with condensation domes, and different discharge slots crossing at sharp angles; E. Replacement pockets along the wall (Photos J.-Y. Bigot).

The discharge slots, described above, are exclusive to this type of caves. The sulfide-rich waters can form pools or a sheet flow corroding the edges and the floor of the passages. This creates the typical almost horizontal pavement of these caves (Figs. 2D, 3A, and 4D), and corrosion notches with a flat roof carved by the aggressive waters. Above the water level condensation-corrosion processes dominate creating a set of typically rounded and smoothed morphologies such as wall niches (Figs. 2A-C, 3C, and 4D-E), ceiling cupolas (Figs. 2D, 3D, and 4C), ceiling channels and megascallops (Fig. 4A). These last are similar in form to scallops created in phreatic conditions, but are entirely related to the flow of air and the accompanying condensation-corrosion processes. Above the slots discharging thermal sulfidic waters these condensation processes can be very strong and cre-

Figure 3. Acqua Fitusa cave: A. Corrosion table and discharge slot, with lateral convection niches; B. Flat ceiling with fretted surface and replacement pockets; C. Replacement gypsum still occupying replacement pockets and convection niches; D. Ceiling cupolas and ceiling dome (Photos Marco Vattano).

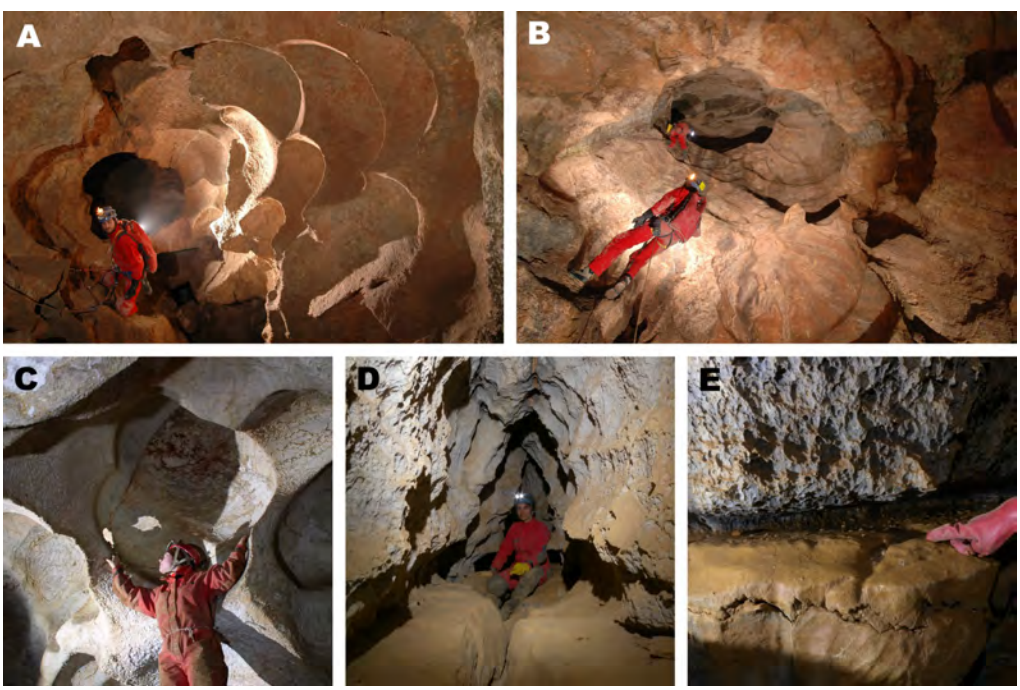

Figure 4. Kraushöhle cave: A. mega scallops in vertical chimney; B. Technical climbing of a chimney revealed that it terminates completely blind; C. Portal connecting two adjacent cupolas; Bad Deutsch Altenburg caves: D. Fault controlled corrosion table with discharge slot (note the similarity to Fig. 2A and 3A); E. Detail of the water table notch with smooth surface below, and wall sculptured by replacement pockets above the former water table (glove for scale) (Photos Lukas Plan).

Table 1 – Typical sulfuric acid cave morphologies in the four described caves.

Cave Morphology	Grotte du Chat	Acqua Fitusa Cave	Bad Deutsch Altenburg caves	Kraushöhle
Maze cave	x	x	x	x
Elongated anastomotic passage				x
Discharge slots	x	x	x	
Sulfuric acid chimney	x			x
Sulfuric notches with flat roof		x		x
Wall convection niches	x	x	x	x
Ceiling cupolas	x	x		x
Condensation-corrosion channels	x	x		x
Megascallops		x		x
Condensation domes	x	x		x
Boxwork	x	x	x	x
Weathered walls	x	x	x	x
Replacement pockets	x	x	x	x
Sulfuric karren				x
Triple junction drip holes				x
Sulfuric cups				x
Corrosion tables	x	x	x	x
Replacement gypsum crusts	x	x		x
Calcite popcorn		x		

ate dome-shaped rooms (Figs. 2D and 3D). Condensating waters corrode the walls deeply, putting in relief more resistant veins (boxwork) (Fig. 2B) and fossils. One of the most characteristic forms are the replacement pockets, small centimeter-sized hollows sometimes still maintaining the replacement gypsum that is at the origin of their formation (Figs. 2C, 3C, and 4D) (De Waele et al., 2009; Plan et al., 2012). Acid dripwaters can also create dissolution hollows and dripholes on the floor, while sulfuric acid trickles carve the rock in karren. Table 1 summarizes the morphologies present in each of the four studied caves. Figures 2-4 show some of the typical sulfuric acid morphologies.

CONCLUSION

Sulfuric acid water table caves are among the most characteristic hypogenic caves in limestone, displaying a typical set of morphological and mineralogical characteristics that are often easy to recognize. Their evolution is strictly related to the base level position and since they have a very fast evolution they are very precise records of base level changes. Some of the minerals typically present in these caves are formed during the speleogenetic phases and their dating precisely marks the age of cave formation, and thus of base level position.

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MORPHOLOGICAL EFFECTS OF CONDENSATION-CORROSION SPELEOGENESIS AT DEVILS HOLE RIDGE, NEVADA

Y.V. Dublyansky[1] and C. Spötl[1]

The Devils Hole Ridge, a small block of Paleozoic carbonate rocks surrounded by the Amargosa Desert in southern Nevada, is located at the discharge end of the Ash Meadows regional groundwater flow system.

Continuous, long-term presence of slightly thermal (33.6°C) groundwater and the extensional tectonic setting, creating underground thermal lakes in open fractures, lead to intense dissolution above the water table. The morphology of the subaerial parts of the tectonic caves was slightly modified by condensation corrosion, and the Devils Hole Prospect Cave was almost entirely created by condensation corrosion. Caves and cavities in the Devils Hole Ridge are an interesting example of a hypogene speleogenesis by mechanism by condensation corrosion, operating above an aquifer which was demonstrably supersaturated with respect to calcite for hundreds of thousands of years.

STUDY AREA

Devils Hole Ridge is a small elongate limestone mountain jutting out about 250 m above the floor of the southern Amargosa Desert — a ca. 200 km^2 intermontane basin extending about 85 km along the California-Nevada border. At the south-west foothill of the Devils Hole Ridge lies the Ash Meadows Oasis, the discharge area of the regional-scale Ash Meadows Groundwater Flow System (Fig. 1).

Limestone of the Cambrian Bonanza King Formation exposed in the Devils Hole Ridge represents the lower carbonate aquifer underlying the Ash Meadows groundwater basin. The aquifer is primarily recharged by infiltration of snowmelt and rainfall at the upper elevations of the Spring Mountains (~3600 m a.s.l.; Winograd et al., 1998). The aquifer is alternately confined by young and partly indurated sediments beneath valleys and unconfined beneath ridges. Flow in the aquifer is little affected by

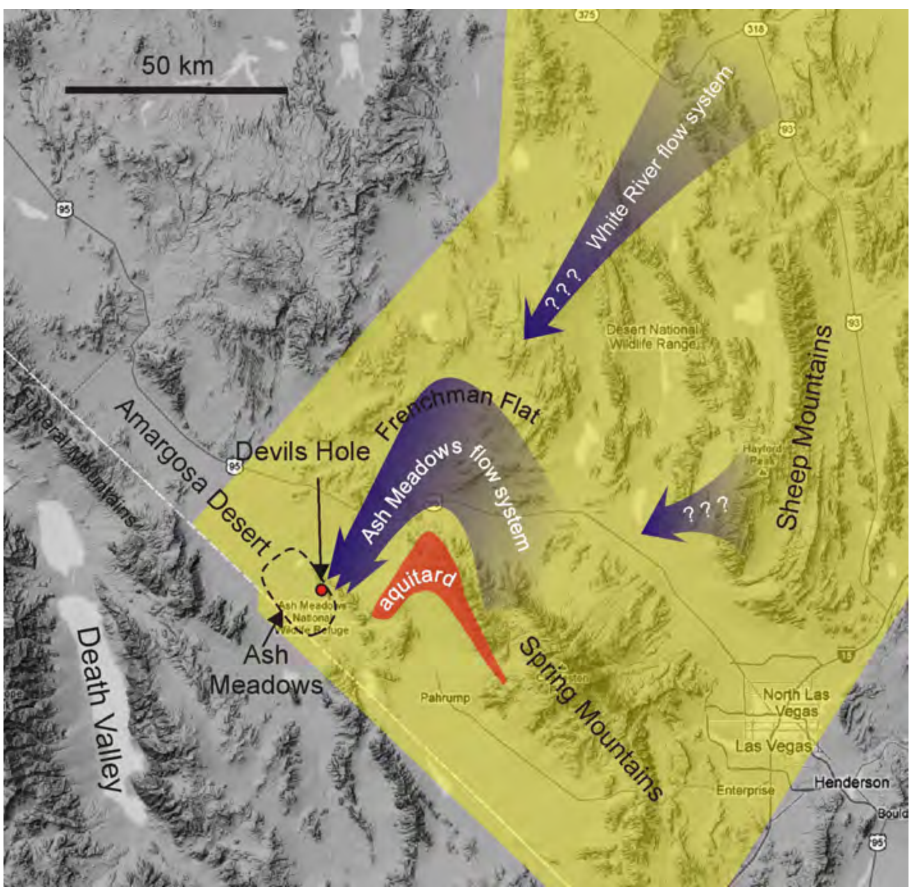

Figure 1. The Ash Meadows Groundwater Flow System. The location of the carbonate corridor (yellow) and the main flowpath of the Ash Meadows Flow System are redrawn from Riggs and Deacon (2002).

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surface topography or drainage. The most conductive flow paths are associated with the NE-SW-oriented fractures, which were opening due to regional extension which started 10 to 5 myr ago (Carr, 1984; Riggs et al., 1994).

At Ash Meadows Oasis, water discharges along a linear array of springs that emerge from flat-lying Quaternary lake beds and local travertine deposits. Water discharging at Ash Meadow is slightly supersaturated with respect to calcite (SI = 0.2; Plummer et al., 2000). The discharging water is moderately thermal, with temperatures up to 32°C in Ash Meadows springs and showing a constant value of 33.6°C in the aquifer intersected by Devils Hole and Devils Hole #2 caves (Coplen, 2007; this study). These temperatures are 13.0 to 14.6°C higher than the local mean annual air temperature (Winograd and Thordarson, 1975).

The Amargosa Desert at Ash Meadows contains abundant and thick calcareous and siliceous spring and marsh deposits of the Pliocene age (2.1 to 3.2 Ma; Hay et al., 1986); groundwater-deposited calcite veins of Pleistocene age are present in alluvium and colluvium in the area (500 to 900 ka; Winograd and Szabo, 1988). The Devils Hole, located at the foothills of the Devils Hole Ridge contains a thick crust of mammillary calcite, which was depositing subaqueously over the last 500 kyr (Ludwig et al., 1992; Winograd et al., 1992). Taken together, available data indicate that groundwater at the discharge end of the Ash Meadows Groundwater Flow System has always been slightly supersaturated with respect to calcite for at least the last 3 myr.

The Devils Hole Ridge has a surface area of ca. 2 km^2 and its vadose zone is up to 250 m thick. Because of the very small potential recharge area and arid climate, active epigenic (supergene) speleogenesis is not expected in the vadose zone of this ridge, even under climate conditions of higher humidity in the past. Yet, we identified ample evidence of specific solution forms in the carbonate rocks of the Devils Hole Ridge.

CAVES AND CARBONATE DISSOLUTION IN DEVILS HOLE RIDGE

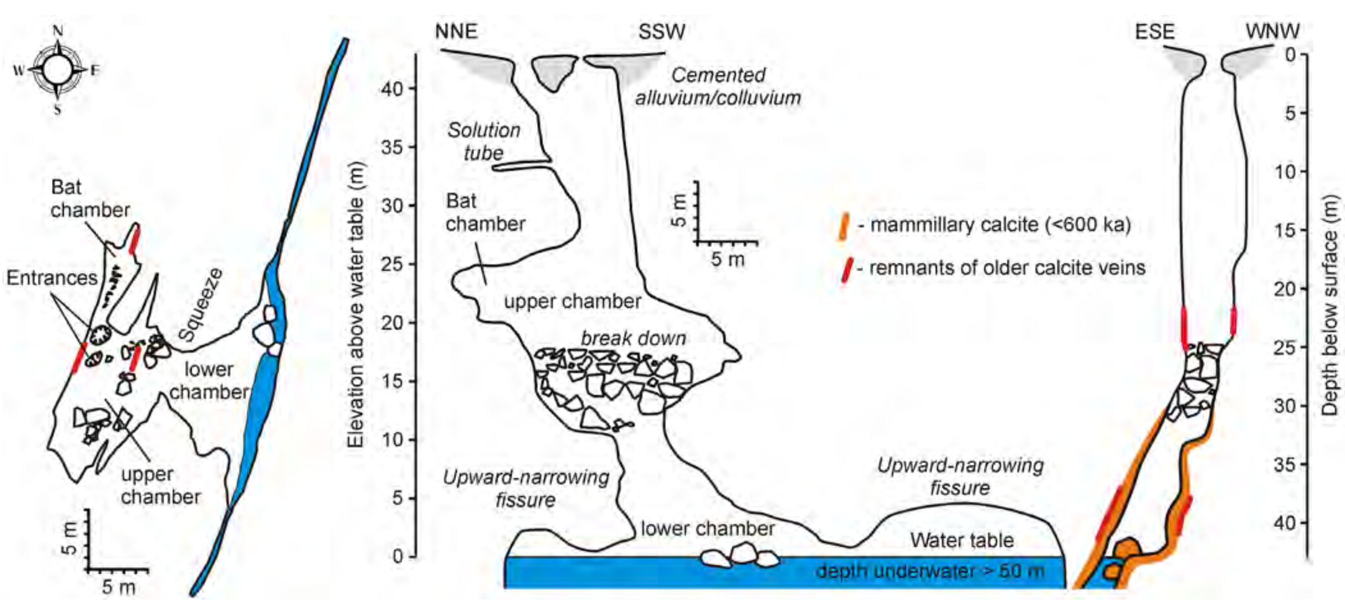
The two major caves in the Devils Hole Ridge are Devils Hole and Devils Hole #2. Although initially the origin of Devils Hole was attributed to solution enlargement of a tectonic fracture (Winograd and Thordarson, 1975), subsequently, Riggs et al. (1994) presented a compelling set of arguments supporting a purely tectonic, extensional origin of the cave. Dissolution due to condensation corrosion was first reported by Riggs and Deacon (2002) for Browns Room – an aphotic chamber located ca. 50 m to the north of the main entrance (collapse) of Devils Hole. Condensation corrosion appeared to be a minor speleogenetic mechanism, only capable of destroying secondary carbonate formations and modifying the cave wall surface.

Working in the area since 2010, we have commonly observed solutional forms in Paleozoic carbonate rocks of the Devils Hole Ridge.

Solutional forms in Devils Hole #2

Devils Hole #2 is a NE-striking, extensional fissure, which intersects the water table at a depth of 43 m (Fig. 2). It is an almost complete analog of the Devils Hole in terms of its (tectonic) origin. The fissure was concealed at the surface by a ca. 1.5 m-thick layer of carbonate-cemented alluvium/colluvium of Pliocene-Pleistocene age. The cave was opened when this indurated layer was breached by condensation corrosion. The cave consists of two chambers separated by a "plug" of collapsed blocks. The upper chamber is near-vertical, 5 x 15 m-wide and 25 m-high. Two calcite veins up to 40 cm-wide are present on the eastern and western walls of the cave; the veins filled two small-aperture extensional fractures, which served as precursors of Devils Hole #2 proper. The latter formed when the bedrock

Figure 2. Plan view (left) and projections (right) of Devils Hole #2 cave.



blocks located between these precursor fractures collapsed. A crawlway between boulders leads to a ledge at the top of the inclined (ca. 60°) lower chamber. Both the hanging- and footwall in this chamber are coated with carbonate deposits, which reach up to 125 cm in thickness.

Solution-smoothed walls — In the upper chamber, the walls are coated with mammillary calcite, up to a height of ca. 2 m above the chamber's floor (ca. 23 m above the water table). The coating is visibly thinned-out by corrosion, so that its initial thickness cannot be determined. The limestone walls above the calcite coating are also smoothed by dissolution (Fig. 3).

Flat channels along fractures and calcite veins — Small channels with characteristic solution morphology, precluding any involvement of gravity-driven water flow, cut through both the bedrock limestone and the old calcite veins on the eastern and western walls of the upper chamber.

Cupolas in indurated alluvium-colluvium — In the ceiling of the upper chamber, composed of carbonate-cemented alluvium-colluvium, dissolutional cupola are present. The solution surfaces cut uniformly across massive limestone bedrock as well as alluvium-colluvium consisting of cobbles, finer-grained material, and carbonate cement. The cupolas are locally separated by millimeter-thin partitions.

"Punk rock" — The southern wall of the upper chamber shows a layer of soft, powdery material, extending into the wall up to 2 cm and eventually grading into hard, unaltered limestone. Despite its softness, the layer preserves the original structure and coloration of the bedrock. This alteration zone is similar to the "punk rock" described by Hill (1987) from New Mexico caves.

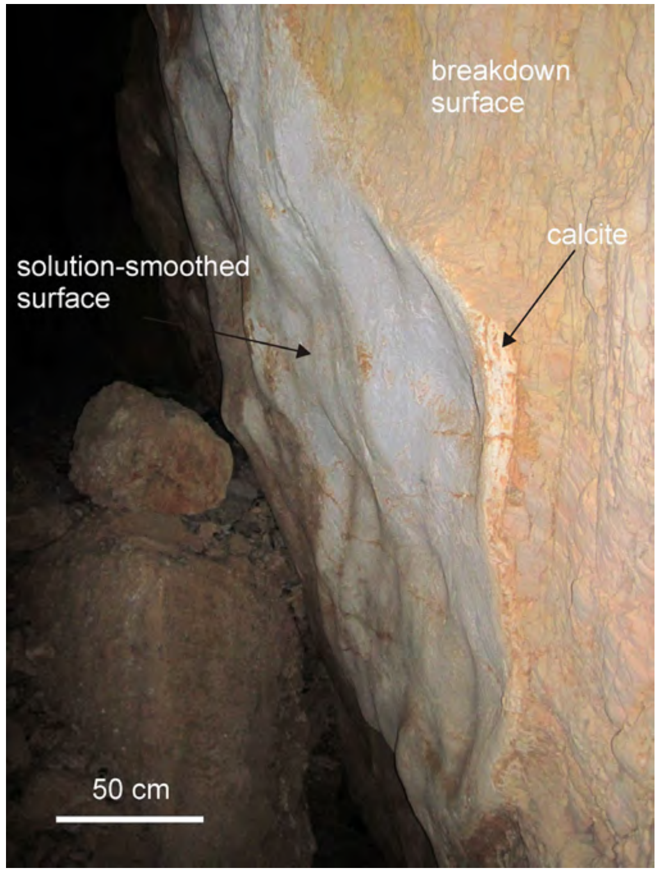


Figure 3. Solution-smoothed wall in the upper chamber of Devils Hole #2. Note that a solution surface cuts uniformly both mammillary calcite and hostrock.

Solutional Forms in Devils Hole Prospect Cave

The cave is located 50 m north of and ca. 12 m higher than Devils Hole (Fig. 4). The entrance to the cave was opened during mineral prospecting work, in the course of which indurated alluvium-colluvium material filling a ca. 1.5 m-wide fracture in the limestone was removed. Unlike other caves in the Devils Hole Ridge, whose internal volume was created primarily by extensional tectonics, the Devils Hole Prospect Cave owes its existence almost exclusively to dissolution. The cave is carved in both Paleozoic limestone and in indurated fault filling. Parts of the cave hosted by carbonate rocks show a characteristic solution-related morphology with abundant cusps, pendants, partitions, spherical niches, and cupolas (Fig. 5). Similar morphologies can be seen in the fault-fill material, but because of the poorer mechanical stability of this material, they are less well preserved.

Solution Cavities at the Topographic Surface

We found numerous small-scale solution cavities on the slopes of the Devils Hole Ridge, within 18 to 41 m-elevation range above the present-day water table. Cavities are exposed on bedrock slopes, and are truncated by solutional slope retreat to different degrees: some of them open as small-diameter (7-10 cm) holes which widen inward, whereas others are severely truncated or even almost completely destroyed (Fig. 6). The common morphological feature of these cavities is their circular cross-section in plan view; in vertical projection they show spherical, spheroidal, and tubular shapes. The maximum diameter of the cavities ranges between 10 and 130 cm. Most of the cavities have barren walls, but some host speleothems forming in subaerial conditions (corallites, popcorn).

MICROCLIMATE AND EVIDENCE OF ACTIVE CONDENSATION CORROSION IN DEVILS HOLE #2

Devils Hole #2 is the only cave in the study area hosting a thermal lake and only partly open to the surface. A 24 hour-monitoring on November 8-9, 2011, demonstrated that cool, dense, low-humidity outside air descends and displaces the less dense, humid cave air, penetrating all the way down to the bottom of the lower chamber. The RH in 1 m above the thermal water table

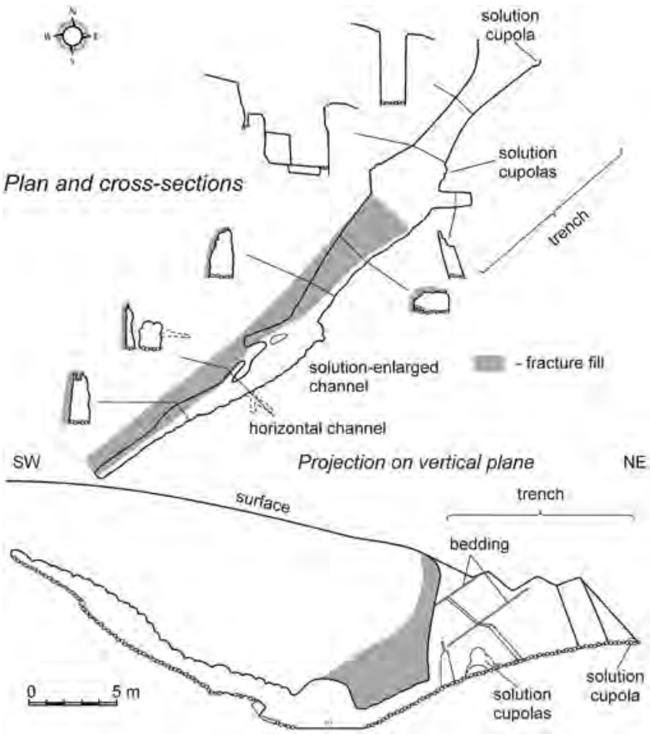


Figure 4. Plan view and projection of the Devils Hole Prospect Cave.

ranges between 60 and 70 % (Fig. 7).

The long-term monitoring (Nov 2008 to Oct 2010) in two chambers of Devils Hole #2 showed that the RH remained low (40 to 70 %) from October through April. From May on, as the outside air became warmer than the air in the cave, invasions of outside air became less frequent and the RH in the lower chamber gradually increased, stabilizing for a short period at 90-94 % (July). It is possible that during this period (two to three weeks during the year) the RH in the lowermost (closest to the thermal lake) part of the lower chamber exceeds 100%, enabling condensation. The data suggest that condensation corrosion is presently not an active speleogenetic process in Devils Hole #2. It was observed only in a small semi-isolated, chamber just above the water (Fig. 8). The walls of the chamber host patches of white powdery material, as well as droplets of water. This material is the calcite precipitate, which forms when the film of condensed water dries out. In contrast to the hanging walls of the lower chamber, coated by dense folia, the walls of this small chamber (at the same elevation above water) are smooth due to solution.

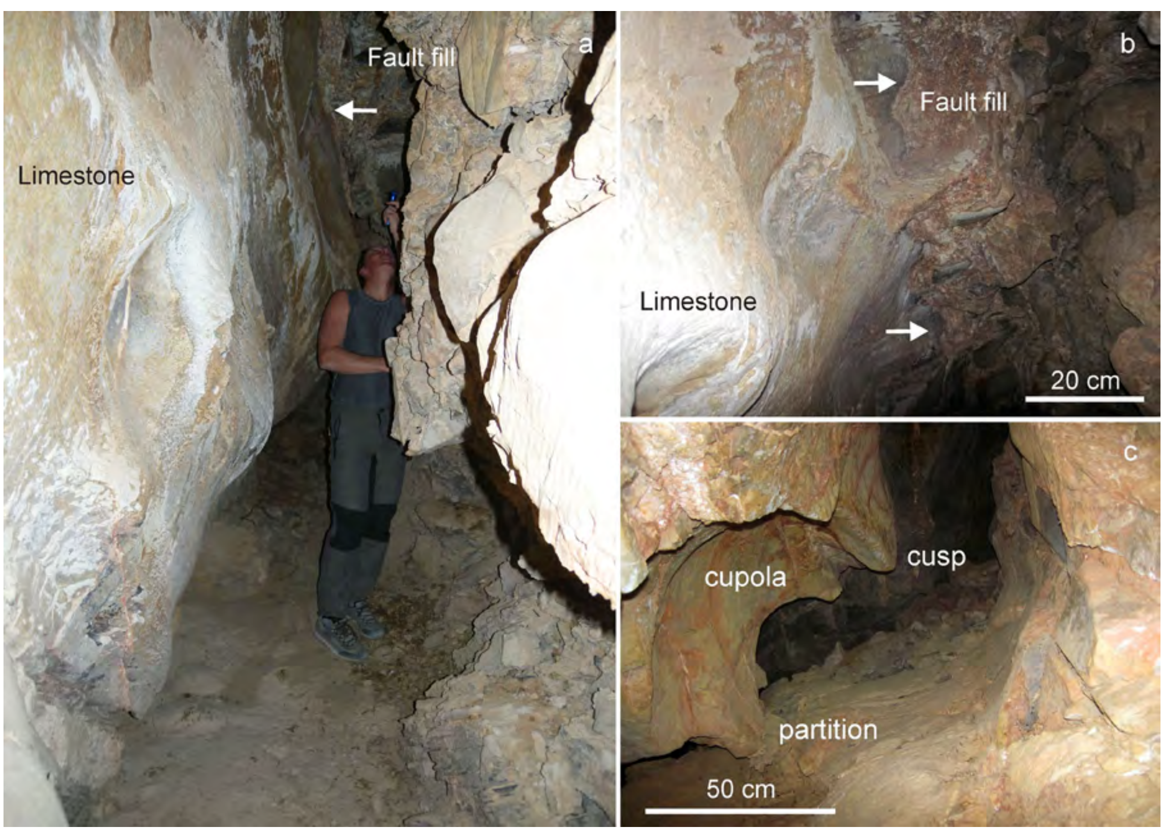


Figure 5. Solutional morphology of the Devils Hole Prospect Cave: a and b – rift-shaped passage with left wall in limestone and right wall in fault fill (arrows show contact between limestone and fault fill); c – cupola-, partition-, and cusp morphologies in massive limestone.

Figure 6. Solutional cavities on the southwestern slope of Devils Hole Ridge showing different degrees of truncation by slope processes: a–c – barely opened cavities (widen inward); d – combination of a vertical tube and a small cupola cut by slope erosion; e - three cavities (arrows) truncated to different degrees (cavity on the left hosts subaerial calcite speleothem; the other two are barren); f - cavity hosting popcorn-like speleothems. No leading fractures are observable; dissolution cuts across various lithologies (b).

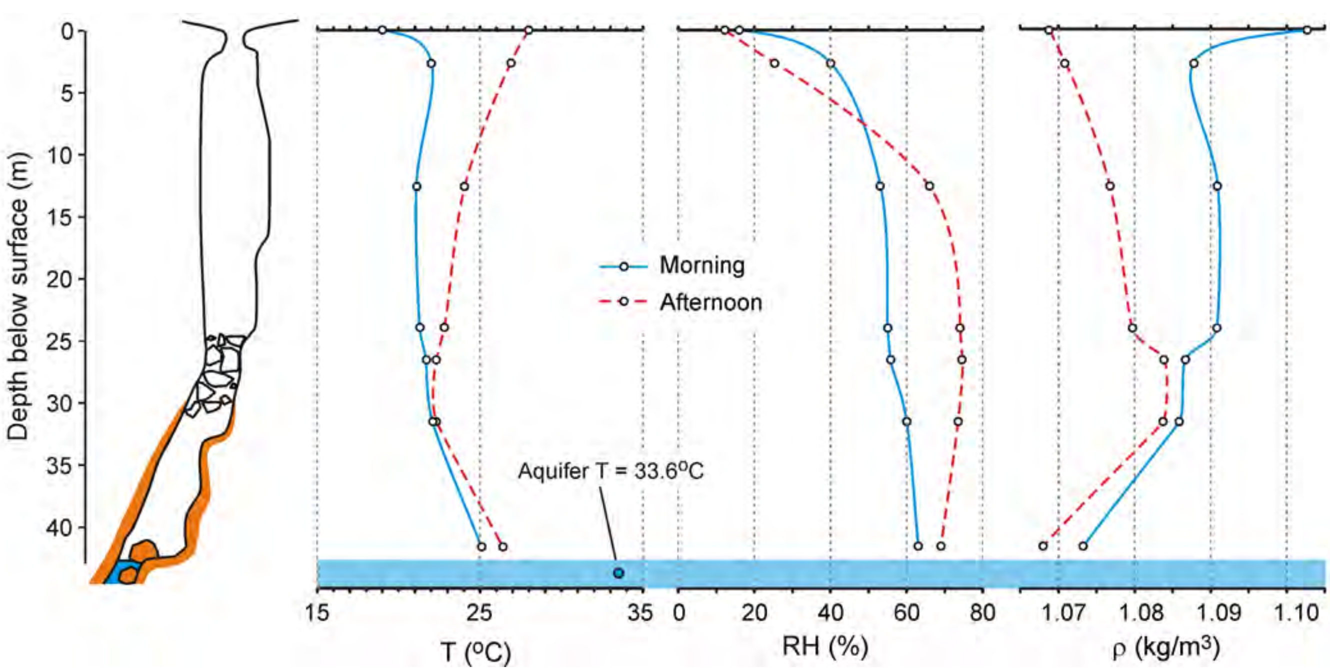


Figure 7. Temperature, relative humidity, and calculated air density in Devils Hole #2 in the morning and in the afternoon.

DISCUSSION: CONDENSATION CORROSION INSIDE DEVILS HOLE RIDGE

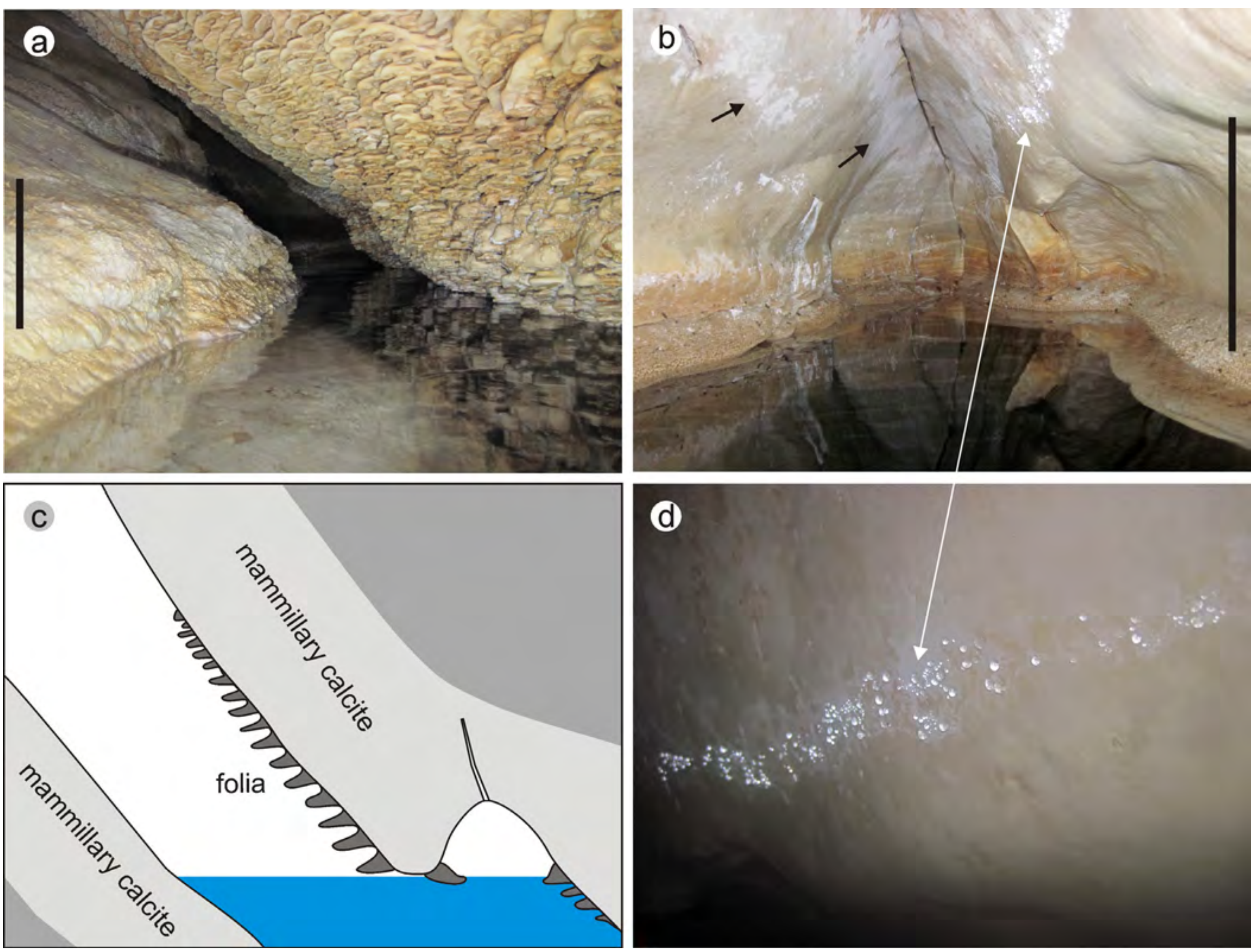
Controls of condensation corrosion

The condensation-corrosion features described above formed when the cavities inside the ridge were essentially closed to the outside atmosphere. For this type of condensation corrosion to become an efficient speleogenetic mechanism, a number of conditions must be met.

Structural conditions — One of the parameters controlling the amount of the evaporation from a thermal water table (i.e., the amount of water available for condensation corrosion), is the surface area of the subterranean thermal lake. In most hypogene caves described so far (e.g., Cigna and Forti 1986; Audra et al., 2002) the process occurred during late stages of hypogene karstification, above thermal lakes. At Devils Hole Ridge, subaqueous karstification can be ruled out, because the water has been saturated with respect to calcite for hundreds of thousands of years. The subterranean voids occupied by the thermal lakes were instead created by extensional tectonics. Because the region has remained in an extensional regime since the mid-Tertiary, such subterranean thermal water surfaces are likely to be widespread.

In addition to creating evaporation surfaces, extensional fractures provide an open space above the thermal water table in which free convection of moist air can develop. Extensional tectonics, collapse, and condensation corrosion operate in concert in creating and enlarging the open space underground. The result of this concerted action may remain unnoticed at the topographic surface.

Figure 8. Active condensation corrosion in a small semi-isolated chamber close to the water table in Devils Hole #2. a – massive folia on the hanging wall in the main part of the cave; b – solution walls in the small chamber (note white powdery material on the walls (black arrows) – the result of dissolution and re-evaporation of water); c – schematic vertical transect showing the position of the small chamber shown in b; and d - condensation droplets on the powdery material in the small chamber. Temperature of the pool water is 33.6°C. Vertical black bar in a and b is 25 cm.



Thermal conditions — A temperature gradient exists between the body of thermal water and the land surface. The aquifer underneath Devils Hole Ridge maintained a nearly constant temperature of ca. 34°C for (at least) hundreds of thousands of years. For most of the late Pleistocene, thermal gradients in the vadose zone were likely greater than they are today because the thermal water table was 6 to 9 m higher (Szabo et al., 1994). Thermal conditions favorable for condensation corrosion may have persisted for an equally long period of time. Extensional-tectonic conditions could also create open fractures which are open to the surface but do not intersect the thermal water table. Inflow of cool air in such fractures provides additional cooling of the rock mass.

Microclimatic conditions — Condensation corrosion is least intensive near the thermal water table, where the difference in temperature between the thermal water pool and the surrounding rock is small. As the warm moist air buoyantly moves up along an open fracture, it cools down, and at certain distance its temperature may reach thermal equilibrium with the surrounding rock. Both buoyant movement and condensation will then no longer be possible. Additionally, the heat released due to condensation warms the rock surface, decreasing the local thermal air-rock gradient. For condensation to proceed, this heat must be removed through solid-state thermal conduction and/or by downward flow of the condensate. It can be inferred, therefore, that there is a certain combination of parameters at which condensation corrosion is most effective; this combination is specific to a given cave configuration.

When a channel or cupola, carving its way up through the rock, opens to the surface, cool dry air flowing into the cave dilutes the ascending moist air. This makes condensation corrosion less efficient, and increasing openness of the cave to the outside atmosphere eventually terminates the process entirely.

Age Constraints for Condensation Corrosion in the Devils Hole Ridge

Solutional channels intersect the mammillary calcite veins in the upper chamber of Devils Hole #2. Although the latter have not been dated radiometrically, they are likely older than 588±39 ka (the oldest age of the main-stage calcite in Devils Hole #2, which overgrows the petrographically and isotopically identical veins in the lower chamber). Thus, the age of ca. 550-600 ka can be considered as a minimum estimate for the initiation of active condensation corrosion in the vadose zone of the Devils Hole Ridge. Before that time the water table could have been significantly higher (Winograd and Szabo, 1988) so that the present-day vadose zone could have been, partly or entirely, under phreatic conditions.

A subaerial speleothem (popcorn) from a condensation corrosion cavity exposed at the surface of the Devils Hole Ridge yielded U-Th ages of 413±12 ka and 318±8 ka. These dates provide a lower limit for the age of this particular cavity. The maximum time span available for the formation of this 40 cm-diameter cavity is thus on the order of 150-200 kyr. Theoretical calculations suggest similar values, e.g. the formation of a spherical cavity with a 50 cm-diameter by condensation corrosion requires 200 to 250 kyr (Dreybrodt et al., 2005). Having started in the Late Pleistocene, the condensation corrosion process inside Devils Hole Ridge locally continues in the vadose zone today, as suggested by observations on condensation in a small semi-isolated chamber at the bottom of Devils Hole #2.

CONCLUSIONS

Condensation corrosion is the only karstic speleogenetic mechanism operating in the Paleozoic limestones of the Devils Hole Ridge since the Pleistocene. It develops above the water table of a slightly thermal aquifer, despite the fact that this groundwater has been saturated with respect to calcite at least for the last ca. 600 kyr. The effects of condensation corrosion range from the modification of the surfaces of underground extensional cavities to the creation of purely solutional cavities. One of the three caves known in the ridge, Devils Hole Prospect Cave, owes its origin almost entirely to condensation corrosion. No sizable underground cavities attributable to epigenic (supergene) karst were found. Our observations confirm the role of condensation corrosion as an important speleogenetic mechanism, capable not only of modifying the pre-existing cave morphology, but also of creating significant caves.

ACKNOWLEDGMENTS

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CONDENSATION CORROSION: MEASUREMENTS AND GEOMORPHIC EVIDENCE IN THE FRASASSI CAVES

Sandro Galdenzi[1], Daniel S. Jones[2,3], and Jennifer L. Macalady[3]

INTRODUCTION

The condensation of acidic waters on subaerial carbonate surfaces (condensation corrosion) can be an important speleogenetic agent under certain conditions (Cigna and Forti, 1986; Sarbu and Lascu, 1997). Specific morphologies associated with condensation corrosion include notches, niches, cupolas, megascallops and domes (Audra, 2009), and have been recognized in many caves from different regions of the world and from different geologic settings. Condensation corrosion can be particularly important in thermal caves, where temperature differences facilitate air convection and water condensation, as well as in sulphidic caves, where degassing and subsequent oxidation of hydrogen sulphide (H_2S) gas provides a ready source of acidity to the subaerial cave environment.

In pioneering studies on the formation of sulphidic caves, condensation corrosion via H_2S degassing and oxidation to sulphuric acid was considered the primary mechanism for speleogenesis (Principi, 1931; Egemeier, 1981). However, recent research has cast doubt on the importance of subaerial H_2S oxidation for sulphidic cave formation (Engel et al., 2004). In the Frasassi cave system, Italy, morphological evidence for both subaerial and subaqueous limestone dissolution has been extensively documented (Galdenzi, 1990; Galdenzi and Maruoka, 2003). In particular, corrosion above the water table has resulted in the formation of massive gypsum deposits as well as specific passage morphologies. Measured rates by Galdenzi et al. (1997) corroborated morphological evidence that condensation corrosion is important at least under certain conditions. Therefore, in order to better define the role of subaerial processes in the Frasassi cave system, we quantified sulphide flux to the cave atmosphere in the modern cave environment, and documented morphological evidence for subaerial corrosion in the past.

SULPHIDE DYNAMICS IN THE MODERN FRASASSI CAVE SYSTEM

The Frasassi caves are organized in several superimposed levels that encompass over 25 km of total passage, mainly in the Jurassic Calcare Massicio Formation. The lowermost levels are directly influenced by the action of H_2S from a sulphidic aquifer. H2S-rich springs emerge at the water table, and form anoxic and microoxic streams and lakes with dissolved sulphide concentrations up to 650 μM.

The Northeast region of the cave is characterized by fast flowing and turbulent sulphidic streams. Where these streams interact with cave air, three different processes act to remove dissolved sulphide from the stream: (i) H_2S degasses from streams to the cave atmosphere (Fig. 1), where it can be oxidized in condensation droplets on the walls and ceiling; (ii) microbially-mediated H_2S oxidation occurs in benthic stream biofilms; and (iii) abiotic sulphide oxidation occurs in the bulk stream water. We quantified each of these processes using field measurements and transport modeling. Complete details and methods are presented in Jones et al. (in review).

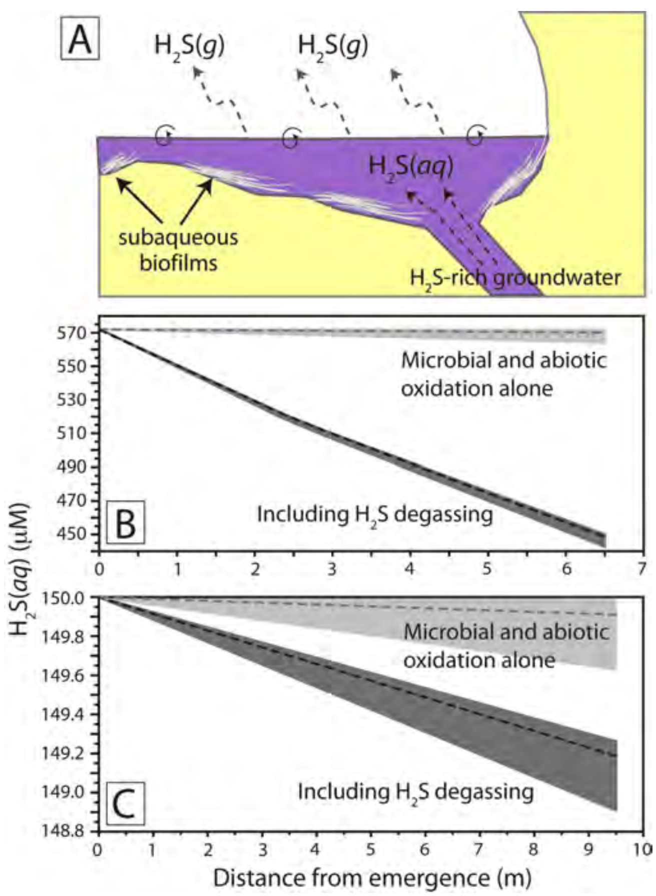

Figure 1. (A) Schematic of a Frasassi cave stream depicting H_2S inflow and degassing. (B and C) Modeled downstream change in H_2S in two streams, with and without including the degassing flux. Shaded areas indicate the range of uncertainties in microbial oxidation rate. After Jones et al. in review.

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We found that H_2S degassing accounted for the majority of sulphide disappearance from Frasassi streams, highlighting the potential importance of subaerial sulphide oxidation for speleogenesis. Model results (Fig. 1) showed that 60-98% of the sulphide lost from streams was due to degassing, and that microbial oxidation below the water table was responsible for the remainder. Abiotic oxidation was largely insignificant. Rapid H_2S degassing near streams is broadly consistent with subaerial corrosion features and the distribution of microbial life above the water table. Near sulphidic streams, cave walls and ceilings are covered with acidic (pH <4) gypsum corrosion residues often >10 cm thick, and the gypsum surface is colonized by extremophilic sulphur-oxidizers that produce highly acidic (pH 0-2) subaerial biofilms (Macalady et al., 2007). Further from flowing streams, gypsum crusts thin and eventually give way to exposed limestone and mildly acidic (pH 6) wall communities (Jones et al., 2008).

GEOMORPHOLOGIC IMPLICATIONS AND EVIDENCE FOR PAST CONDENSATION CORROSION IN FRASASSI

The lower cave levels have a mainly sub-horizontal development, where fully subaqueous passages alternated with free surface zones. The cave is organized as a network of large passages that roughly converge to an emergence point, and are often connected to each other through narrow siphons. Diffuse rims of replacement gypsum as well as corrosion pockets that result from the gypsum removal are testament to the contribution of condensation corrosion above the ancient water table. However, subaerial and subaqueous corrosion features are often overprinted, as passages may be submerged or exposed following changes in the local base level of the Sentino River due to the alternating cycles of erosion and sedimentation in the Sentino River valley. Overprinting can make it challenging to evaluate the true importance of subaqueous versus subaerial processes in past cavern development.

Important morphogenetic effects of condensation corrosion can nevertheless be observed in several exemplary chambers in the Grotta Grande del Vento. Long wide passages with flat floors are produced by sulphidic water rising through subaqueous feeder conduits or directly from narrow slots in the floor (Figure 2a). Downstream, additional sulphidic water recharge occurred through minor infeeder conduits, slots, and crevasses. Above the water table, condensation corrosion acted to enlarge walls and ceiling, producing the final passage section. Some isolated domes developed by the same mechanism, above a main feeder or a system of fissures and crevasses (Figure 2b).

During phases in which the Sentino River progressively downcut into the river valley, the water table in the cave also gradually lowered, and corrosion processes due to sulphidic waters occurred in deeper regions, forming new lower levels. However, old feeder conduits remained and would have provided a pathway for H_2S gas transfer to upper levels. Old feeder conduits

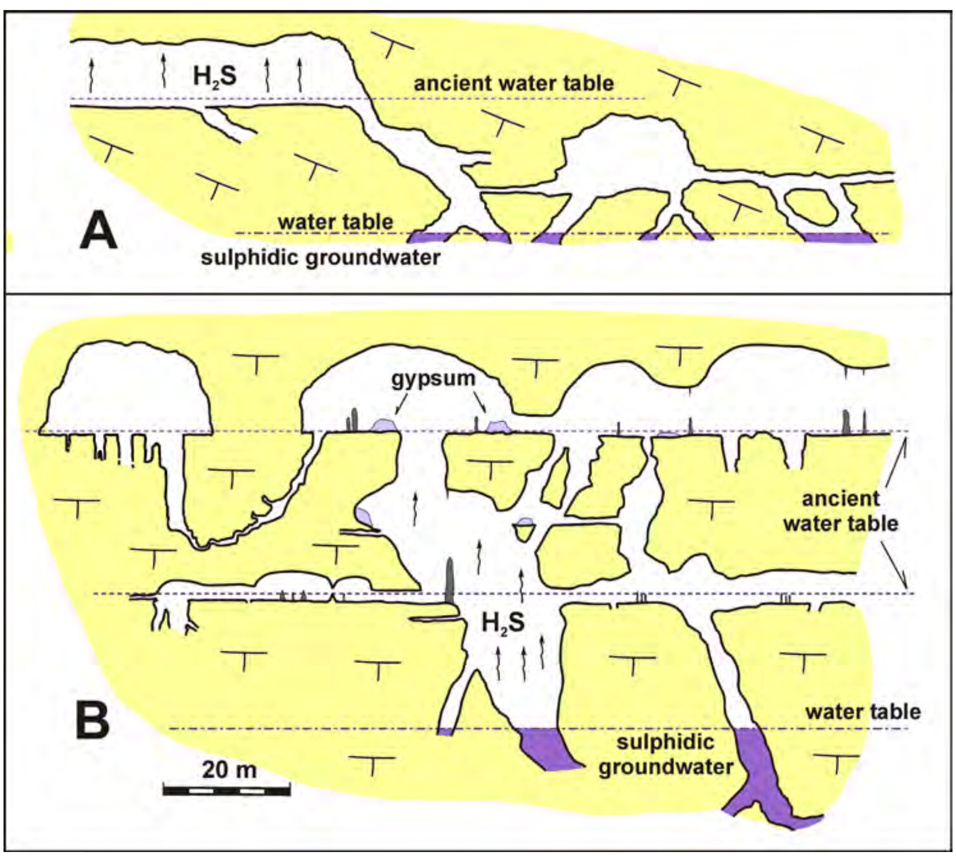

Figure 2. Schematic section of cave rooms enlarged by condensation corrosion. (A) Ancient water-table passages at the outflow of an ascending phreatic passage. (B) Multiple levels produced after successive lowering of the local base level. Domes and passages with flat rock floors developed mainly in the upper level above the past water table.

therefore may have facilitated periodic corrosion and gypsum deposition in the upper dry levels, alternating with the calcite deposition by seepage water.

Widespread relict structures in the upper levels show that degassing and condensation corrosion were of greater importance overall during some periods of the cave history than in the present. This is also confirmed by the size and depositional setting of massive gypsum glaciers in the upper levels, which were derived from H_2S oxidation in the cave air (Galdenzi and Maruoka, 2003) and indicate a level of gypsum production that far exceeds that of the modern sulphidic zones. Differences in the magnitude of degassing throughout the cave history likely resulted from changing hydrodynamic conditions, which can reduce or favor extensive H_2S exchange with the cave atmosphere. It should also be noted that the presence of extensive air-water interfaces would increase the supply of dissolved oxygen to shallow cave waters, which would favor the growth of sulphur-oxidizing bacteria. Chemosynthetic microbial biomass would allow some sulphur re-cycling by sulphate reducers, which could represent a further source of H_2S for degassing and condensation corrosion. Spatial and temporal differences in degassing and condensation corrosion throughout the speleogenetic history of the Frasassi cave system likely contributed to the highly variable sizes of passages and chambers that characterize the contemporary cave morphology.

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PERMIAN HYDROTHERMAL KARST IN KRAKÓW REGION (SOUTHERN POLAND) AND ITS PECULIAR INTERNAL SEDIMENTS

Michał Gradziński[1], Anna Lewandowska[2], Mariusz Paszkowski[3], Marek Duliński[4], Jerzy Nawrocki[5] and Michał Żywiecki[6]

The development of caves influenced by the deep circulation of water has received increasing interest for the last thirty years. Presently, hypogene caves have been recognized all around the world. Conversely, the ancient examples filled with sediments and representing palaeokarst forms are not so common.

The karst forms and their sediment fillings were encountered in the Dębnik Anticline (Kraków region, Southern Poland) composed of Middle Devonian to Mississippian carbonates. The development of karst slightly postdates the Permian (ca. 300 Ma) volcanic activity in the Kraków region. In this region major transcontinental strike and slip Hamburg-Kraków-Dobruja fault zone induced a series of minor, en echelon, extensional faults, which served as magma passages and guided karst conduits.

The karst forms in the Dębnik Anticline reach several to tens of meters in size. They are filled with: i) massive, subaqueous, coarse crystalline calcite spar; ii) crystalloclastic, bedded limestones; iii) jasper lenses; iv) kaolinitised tuffs. The sediments are characterized by red colouration caused by iron compounds.

Coarse crystalline calcite spar composes beds up to several dozen centimeters in thickness. They are laminated and comprise frutexites type structures. The calcites are interbedded with pinkish-red crystalloclastic limestones, which are built of detritic calcite crystals from silt size to a few millimeters across. Some of the crystals are of skeletal type. Crystalloclastic limestones are normally graded. Both calcite spar and crystalloclastic limestones underwent synsedimentary deformations, which resulted in brecciation and plastic deformations.

The above deposits fill karst forms up to a few metres in lateral extent. However, analogously filled enormously huge (up to around 100 m across) forms were recognized in the early 80s of the last century. Presently, they are completely exploited.

The karst forms were fragments of extensive circulation system. It was fed by waters of elevated temperature, rich in endogenic CO_2, which is proved by fluid inclusion analysis and stable isotope investigation. The origin of this system was associated with volcanic activity. The roots of the system are represented by fissures filled with coarse crystalline, red and white calcites of onyx type, which are common in the Dębnik Anticline. Water issuing from this system on the surface caused precipitation of red travertines. These travertines are preserved only as clasts in the Lower Permian conglomerates deposited in the local tectonic depressions.

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HYPOGENE LIMESTONE CAVES IN GERMANY: GEOCHEMICAL BACKGROUND AND REGIONALITY

Stephan Kempe[1]

Germany exhibits a very diverse geological history. Thus, a large number of stratigraphically, petrographically and tectonically different carbonate and sulfate rocks exist that have been subject to karstification. Here, I discuss first the possible "agents" (sensu Klimchouk) of hypogene karstification. Three principally different processes are identified: water rising because of buoyancy (either thermally or concentration induced), in-situ oxidation of siderite, or rising gases (CO_2, CH_4 or H_2S). Next, a rough overview of German caves and karst is presented. If applying the most pertinent epigene versus hypogene morphological characteristics, it becomes evident that hypogene caves occur in many different areas, often side-by-side with clearly epigene caves. For many areas, the agents of hypogene speleogenesis must remain unclear. This applies for most caves in the Paleozoic limestones of the Rhenish Schist Massif. Only the Iberg/Harz caves seem to be a clear case, with the world-wide highest concentrations of siderite weathering-induced caves occur. The large cavities discovered recently in the Blauhöhlen System and some of the deep pit caves in the Swabian Alb may have their explanation in volcanic CO_2, having emanated from some of the 355 pipes of the Swabian volcanic field. Most striking is the high concentration of hypogene caves in the Franconian Alb. Many of them occur in a small area while other areas are devoid of larger caves. Here the tectonic situation suggests that fractures could have taped reservoirs of either sulfide or methane from below. The finding of goethitic crusts in the Bismarckgrotte may indicate that rising anaerobic gases could have been involved.

INTRODUCTION

Hypogene karstification is a more common karstification process than anticipated even 20 years ago. It also appears to be a much more complicated process, involving a geochemistry far beyond that of the "simple" reaction of carbonic acid with carbonate rocks. Hypogene karstification is defined as upward movement "of the cave-forming agent" (sensu Klimchouk, 2007, 2012). This "agent" may be either the water itself rising or a compound available in situ at depth or rising from the depth of the crust. Water can rise either by thermal convection or by natural convection driven by dissolution of an easily dissolvable compound. This latter case is illustrated by some of the German gypsum caves as reported by Kempe (this volume). On the other hand a number of chemical reactions are discussed that give rise to hypogene karstification cause by in situ processes or advection of certain compounds.

In 1971, the author showed that siderite weathering can produce appreciable caves far below the water level (Kempe, 1971, 1975, 1998, 2008, 2009; Svensson and Kempe, 1989).

According to:

$$4FeCO_3 + O_2 + 6H_2O \rightarrow 4FeO(OH) + 4HCO_3^- + 4H^+ \rightarrow 4FeO(OH) + 4H_2O + 4CO_2 \quad (1)$$

The ferrous carbonate, the mineral siderite, is oxidized to the ferric oxy-hydroxide, the mineral goethite, thereby liberating protons and/or free CO_2. These can in turn attack surrounding carbonate rocks, forming hypogene cavities around and near the oxidizing iron ore lode.

The simplest case involving a rising gaseous agent would be CO_2 derived from either volcanic vents or dikes or from other sources at depth. In such a scenario, the caves should be largest very near to the base of the karstified limestone stratum.

In the early 1980s, another process was identified producing hypogene cavities, i.e. oxidation of rising H_2S (e.g., Palmer and Hill, 2012):

$$H_2S + 2O_2 \rightarrow 2H+ + SO_4^{2-} \leftrightarrow H^+ + HSO_4^- \quad (2)$$

This process appears to be capable of corroding very large cavities in the periphery of oilfields, such as in the Captain Reef complex at the Permian Basin, New Mexico and Texas. There some of the largest rooms and the largest caves of the world are found, such as Lechuguilla and Carlsbad Cavern among others (e.g., Kambesis, 2012). The process has been identified as in fact operative because in the process of limestone dissolution, gypsum is forced from solution. In spite of the high solubility of gypsum, it has survived attack by later seepage water in drier corners of the caves, thus forming "the smoking gun" of this important cave-forming process.

However, there may be yet a third process, i.e., methane oxidation (Kempe, 2008):

$$CH_4 + O_2 \rightarrow 4H^+ + CO_2 \quad (3)$$

Since methane is much more common than H_2S in the subsurface, it may play a much larger role than the former in corroding large subterraneous cavities. There is one problem with methane; its oxidation does not leave a "smoking gun." Thus it may be difficult to finally prove its importance. One possibility does, however, exist. If carbonates should be co-precipitated at sur-

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faces that allow degassing, then these carbonates should have a much lower $\delta^{13}C$ signature. This is because crustal methane has a very low $\delta^{13}C$, i.e. up to 70 per-mil lower than the organic matter it has been derived from (Rosenfeld and Silverman, 1959). We all should look out for such cases.

HYPOGENE CAVES OF GERMANY

Germany has a complex geology featuring large tracts of Devonian, Carboniferous, Permian, Triassic, Jurassic and Cretaceous carbonates and sulfates. The Devonian and Carboniferous rocks have been subject to the Variscian orogeny with intensive folding. Permian and Mesozoic rocks have been subject to gradual uplift and fracturing and, in the south of Germany, to Alpine orogenic thrusting (Fig. 1). Thus, there is ample opportunity for karstification under various stratigraphic, tectonic, petrographical, morphological, and altitudinal setting. Moreover, these caves have been under severe permafrost conditions during glacial events. Accordingly, there are numerous different conditions under which caves are forming and have formed in the past.

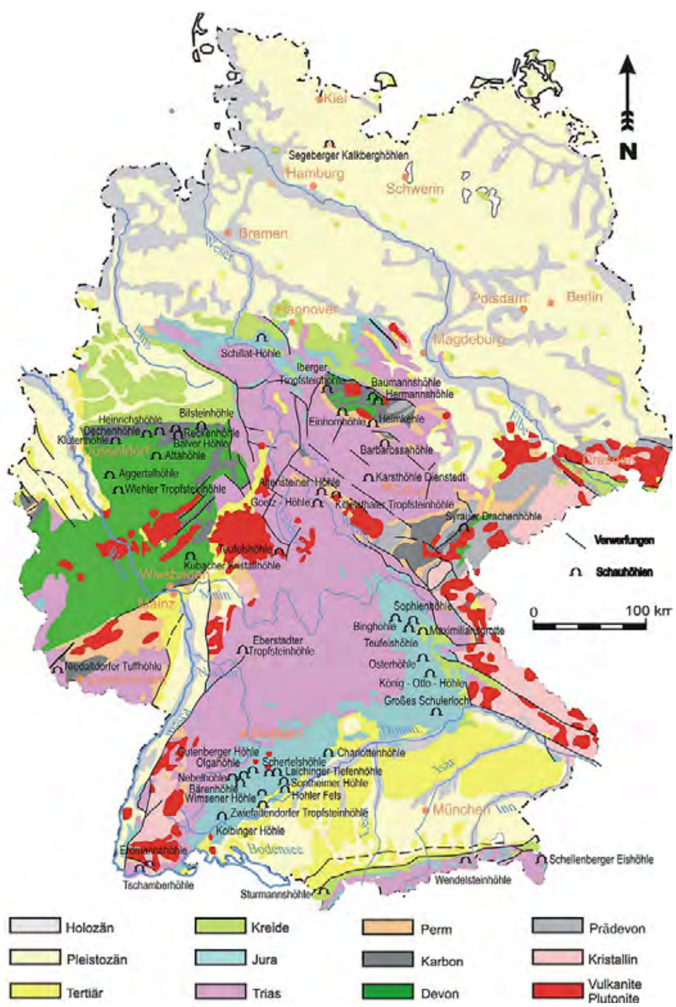

Figure 1. Geological map and show caves of Germany

In spite of the large limestone areas, Germany has only 15 caves longer than 5 km (VdHK, 2013). When looking at cave densities in Germany, we find large differences: Some of the Devonian rocks appear to be Swiss cheese with numerous caves in small areas, while some of the Triassic, Jurassic and Cretaceous rocks appear devoid of significant cavities over large regions. Even if this impression may be skewed because we have yet to find caves in those seemingly cave-less areas, it already suggests that cave formation is not a statistical process, such as one would expect if caves would form exclusively according to "Dreybrodt-type-epigene-cave-models" (e.g., Dreybrodt, 2008).

It is not entirely clear how to separate fossil epigene caves from those of hypogene origin, but Table 1 gives a rough matrix of criteria that may serve for the majority of cases.

According to these criteria, and in some cases without actually having visited the caves, one can try to differentiate German caves (Kempe and Rosendahl, 2008) into the two classes (obviously with leaving room for discussion).

The northern-most cave in Germany is the Segeberger Kalkberghöhle (Fig. 2), a typical, 2 km-long maze cave in the anhydrite/gypsum cap of a salt dome (for more data on sulfate caves, see Table 1 in Kempe, this volume; for data on limestone caves, see Müller and Wolf, 2013; VdHK, 2013). Most of Table 1 criteria above would put the cave into the hypogene category, i.e. the morphology is lacking epigene characteristic because the cave shows only elements of slow, convective dissolution (sloping side walls, solution ceiling, solution cups etc.). Nevertheless it is not clear if the water came from below, rather it seems to have infiltrated from the glacial sediments adjacent to the anhydrite and the altitude of this water exchange was not only near the groundwater level but also was restricted to a thickness of less than 2 to 3 m. Fresh water must have come in along the ceiling of the passages (where they are widened fastest) and left, after being saturated with gypsum, either through floor cracks or laterally back to the glacial sand or till body.

Further to the south, in the upper Jurassic (Korallenoolith) of the Weserbergland, we find a group of caves (Schillathöhle, Riesenbergsystem) that are linear in ground plan but lack other epigene morphological characteristics. Thus, the above table does not allow a clear grouping of these caves that at best are shallow phreatic caves with no real signs of an "agent coming from below.".

To the east of the Weserbergland the Harz Mountains rise, exposing two upper Devonian reef limestone complexes, the Iberg-Winterberg near Bad Grund and the Elbingeroder massif. The Iberg/Winterberg, roughly 2 km^2 in size is a prime example of high density of caves of both groups, hypogene and epigene. The Winterberg, a large limestone quarry, exposed a number of epigene caves: large, steep canyons such as the Mammut Höhle (1.04 km long). Most these are now destroyed, except for the Neue Winterberghöhle. In contrary to this, the neighboring Iberg is the world-wide most important example for hypogene cavities

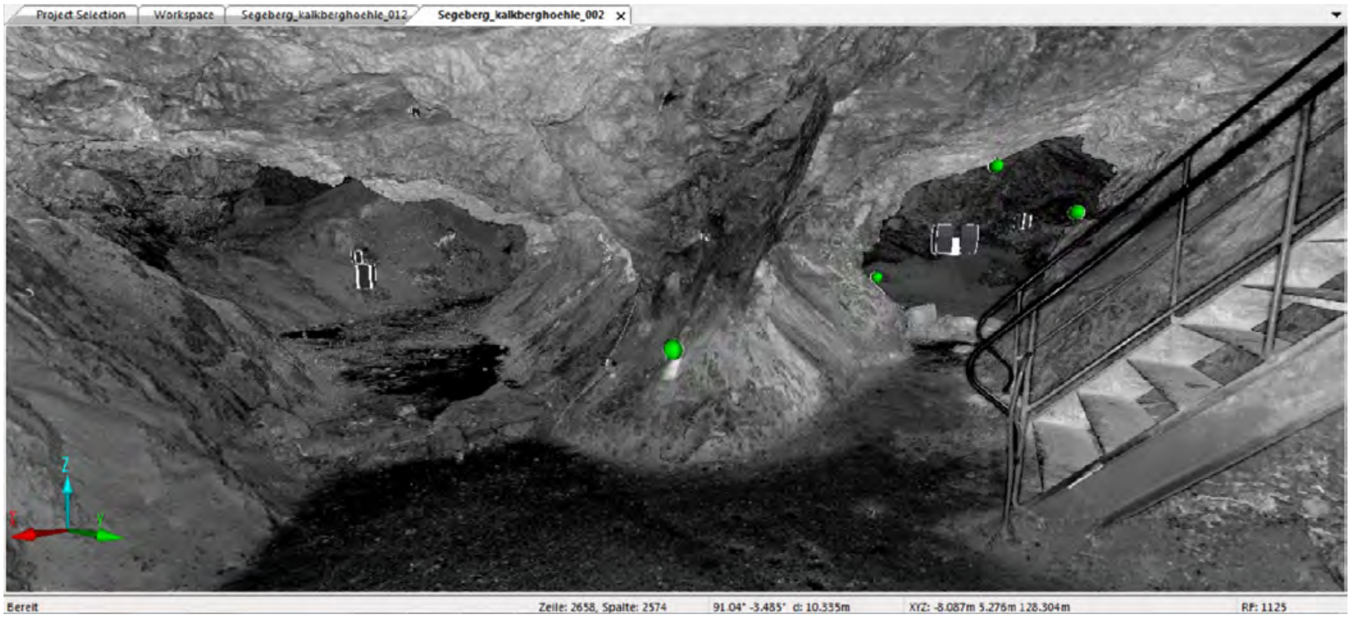

Figure 2. 3D scan of a solutional passage of the Segeberger Kalkberghöhle in anhydrite with typical sloping sidewalls and horizontal ceilings characteristic of a cave created by internal convection.

caused by siderite weathering. Within a square of 250 m, at least 8 km of cavities are recorded (Fig. 3). Iron ore mining has connected these originally isolated caves (Fig. 4) into one large 3D-maze, composed of large irregular halls and passages following former siderite veins (Reimer, 1990). In some ponds in the mine, siderite weathering is still progressing, forming small analogs of this sort of hypogene karstification (Svensson, 1989; Svensson and Kempe, 1989).

The Elbingeroder reef complex is also mined, but no large cave appears to have been discovered. Rather, the caves in Elbingerode concentrate on both flanks of the Bode valley, the Baumannshöhle (1.95 km long), the earliest regularly managed show cave, is located on the west and the longer Hermannshöhle (3.36 km long) on the east. The Baumannshöhle appears to be deep phreatic; none of the epigene characteristics are present. What sort of agent was responsible for its formation remains everybody's guess. The Hermannshöhle on the other hand, is an epigene cave, linear in shape and with an active creek at its lower level. Other caves in the vicinity show solution ceilings and are also phreatic. They are not, however, associated with known iron ore veins.

The Harz and its smaller sister, the Kyffhäuser, are accompanied by upper Permian evaporative rocks along their southern borders containing a score of gypsum/anhydrite caves, many of those are hypogene (see Kempe, this volume). Only one larger dolomite cave is known, the Einhornhöhle (Unicorn Cave), famous for its cave bear bones (Paul and Vladi, 2001). Only part of the cavity is visible, most of its volume is buried by sediments, including thick layers of gravel. In spite of this and its general linear development, it is not clear at all what sort of development the cave took.

The Rhenish Schist Massif contains, similar to the Harz, many small limestone outcrops of middle to upper Devonian age that

Table 1. Criteria to differentiate between hypogene and epigene limestone caves according to own experience.

Evidence for	Epigene	Hypogene
General layout of cave	Linear, tributary, cave of substantial longitudinal extent	Maze, isolated cavities, cave passages limited to a small area
Entrances	Former ponors or springs	No natural entrance, or cave opened accidentally by erosion
Shape of rooms	Canyons, waterfall pits, seepage shafts, or phreatic round or oval passages	Large chambers interconnected (if at all) by narrow passages, shafts, passages of rough cross-section
General wall morphology	Meandering passages, erosion pots, scallops	Ceiling cupolas, solutional ceilings, solution cups, sloping side walls
Wall roughness	Smoothed and polished walls	Walls irregular with pockets, harder seams protruding, fossils exposed
Sediments	Allochthonous or autochthonous gravel, sandy material	Fine grained, autochthonous sediments

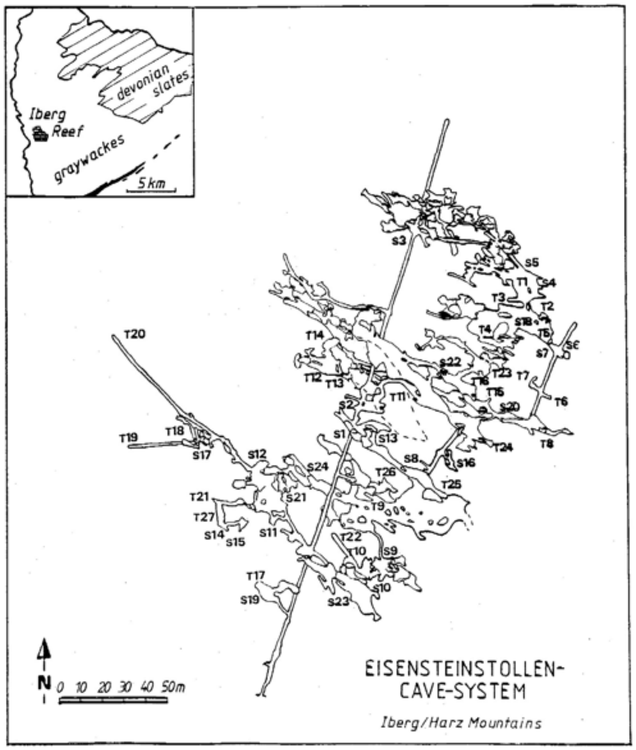

Figure 3. Plan of the 5 km-long Eisensteinstollen, the largest (but not the only) hypogene cavity system caused by siderite weathering at the eastern flank of the Iberg, Harz. It consists of natural large halls connected by mine passages (plan after Reimer and Kempe, 1989, based on survey of Fricke and co-worker). Inset: Location of the Iberg/Winterberg in the northern Harz.

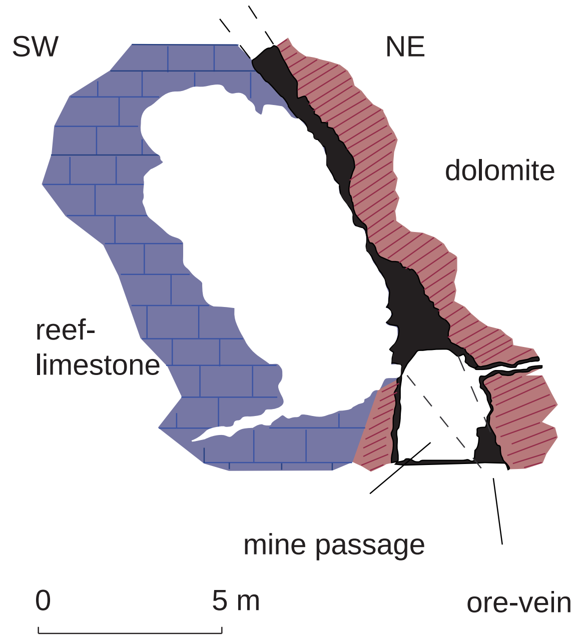

Figure 4. Cross-section through a cavity formed along a former siderite vein and accompanying cave (after Reimer and Kempe, 1989).

show high degrees of karstification. Many of the caves are well-known show caves. Nevertheless, no overview exists about their genesis in general. Some of the caves are epigene, like the newly discovered and still active 6.2 km long Herbstlabyrinth near Herborn (Dorsten and Dorsten, 2012). Others, specifically those that are fossil, are difficult to judge. For example, the Kluterthöhle is a 5.7 km long maze cave (Koch, 1992), suggesting hypogene origin. However, it is also transgressed by epigene waters between a sink and a spring. This could be a secondary stream capture, thus complicating the interpretation. Other maze caves include the Aggertalhöhle and 6.7 km long Attahöhle. Linear caves are the Dechenhöhle and the Heinrichshöhle. At least one cave is clearly of hypogene origin, the Kubacher Kristallhöhle near Weilburg (Fig. 5). It features a nearly 200 m long, up to 23 m wide and 30 m high cavity created by deep phreatic corrosion. Since iron ores have been mined in the vicinity and siderite is a common ore in the Lahn-Dill district, the Kubacher Cave may also have been formed by siderite weathering even though there are not immediate remains of an ore vein in the cave. Also, the largest cave room yet discovered in Germany is situated in the Rhenish Massif and must have been hypogene: the at least 700 m long, 200 m wide and 20 m high cavity at the bottom of the limestone quarry Rhodenhaus-Süd near Wülfrath (Drozdzewski et al., 1998). The cave is entirely filled with quartz sand that contains layers of charred plant remains dating into the Lower Cretaceous. Thus, the cave is at least 100 million years old. Even more, at its center an impressive limonitic ore deposit at least 50 m wide and 10 m thick was unearthed. This suggests that the ascending "cave forming agent" was anaerobic, carrying ferrous iron into the cave and creating a speleogene iron ore deposit possibly similar to what is found in Jordan at Warda (Al-Malabeh et al., 2008) and, to a much smaller extent, in the Bismarckgrotte (see below).

By area, the middle Triassic Muschelkalk formation is one of the largest limestone areas in Germany, occupying much of the center of the country. Nevertheless, it is the limestone formation with the lowest number of caves. The Eberstadter Tropfsteinhöhle, one of the few Muschelkalk show caves, and the parallel, over 3 km-long Hohler Stein are clearly epigenic, however, also of strange character. These caves appear to have been formed by erosion of back-cutting waterfalls and not so much by turbulent dissolution. Other caves in the (upper) Muschelkalk are large maze caves such as the Fuchslabyrinth (9.4 km in length) and the associated Schandtauberhöhle (3.7 km), both today with an epigene hydrology but most likely hypogene in origin (Klimchouk, 2007).

The upper Jurassic (Malm) limestone forms the largest continuous karst area in Germany. It consists of the Franconian Alb in

Figure 5. View into the Kubacher Kristallhöhle, a large cavity formed by convective dissolution as illustrated by the cupolas in the background and the sloping sidewalls (facets) in the foreground.

the east and southeast and the Swabian Alb in the west, divided by the Nördlinger and Steinheimer impact craters in the center (Fig. 1). The south-dipping strata are terminated in the North by an imposing escarpment. Several thousand caves are known in the Malm, however, km-long caves are rare.

In the Swabian Alb, recent exploration by diving and digging down more than 100 m in sinkholes has discovered a large cave system, the 10 km long Blauhöhle (Hinderer and Kücha, 2012). It discharges its water through the Blautopf, one of largest karst springs in Germany. In spite of the fact that the system is largely linear and active, its tremendously large halls do not fit the small tributary area of the present cave nor its missing aggressivity (Kempe et al., 2002). Therefore, doubt remains about what "agent" was responsible in forming this voluminous system. One of the possibilities is that volcanic CO_2 was involved because some of the 355 vents of the Swabian Volcanism dissected the strata 16 to 17 Mio years ago. Their slow degassing could have provided the necessary "agent." If this were true, then a hypogene agent would cause an epigene-looking cave system. In this context, the presence of deep vertical caves such as the 86 m deep Laichinger Tiefenhöhle (1.35 km long) and the 127 m deep Laierhöhle (2.43 km long) may be of interest as well. These are not former sinks of surface waters but may also represent hypogene features (Fig. 6). The Swabian Alb holds many more epigene caves, including the Wulfbachquellhöhle (6.5 km), the Mordloch (4.4 km), the Falkensteiner Höhle (4.0 km) and the Wimsener Höhle, not to mention the large caves system that currently diverts the water of the upper Danube towards the Achtopf and the Rhine. There are also many fossil epigene caves, like the Nebelhöhle and the Charlottenhöhle. Thus overall the character of the larger caves in the Swabian Alb is more epi- than hypogene.

In the Franconian Alb, the general picture seems to be the other way around. Here, clearly hypogene caves dominate. Until the Mühlbachquellhöhle (7.7 km) near the southern border of Franconian Alb was discovered, not many sizeable, epigene caves were known in the area, among them the fossil, linear Bing Höhle and the Fellner Doline, an active, vertical sinkhole. All of the other larger caves (e.g., Kaulich and Schaaf, 1980; FHKF, NHG, 2002) seem to be hypogene. Most of them are formed in the dolomitic stromatolitic sponge "reefs" (Massenkalk), showing irregular large halls which are connected by small passages or pits forming three dimensional mazes. These include the famous Zoolithenhöhle to which William Buckland paid a visit in search of the Deluge (Fig. 7), as well as the show caves Maximiliansgrotte (Fig. 8), Sophienhöhle and Teufelshöhle (1.5 km long). Internationally less well-known is the Geisloch, the northernmost caves of this type with its well-preserved glacially damaged speleothems. Some of these caves are labyrinthic and attain respectable lengths on small areas, such as the Windloch at Kauerheim where 2.2 km of cave fit into 200×100m of area, or the Moggaster Höhle featuring 2 km of passages within an area of 150×100 m and the Bismarckgrotte where we find 1.2 km of passages within an area of 110×100 m. Clearly, hypogene are also the typical maze caves Schönstein-Brunnstein (0.6 km) and Stein-am-Wasser and a score of smaller caves such as the Windloch at Kürmreuth. Some of the caves essentially consist of one large cavity only, such as the Riesenburg and the Rosenmüllerhöhle, named after the scientist who established the cave bear as the first extinct mammal in 1794 and the Oswald-/Witzen-/Wunders- cave system. Others are larger rooms opening below a pit or doline such as the Esper-Höhle, the Breitensteinbäuerin or theWindloch at Großmeinfeld. Many of the even smaller caves (there are about 3000 caves listed with less than 50 m in length) are most likely hypogene but often impacted in their morphology by frost shattering, obliterating their original morphology.

Two clues exist as to their common origin: most of these caves seem to fall into a very narrow NW-SE striking corridor (Fig. 9) and in the Bismarckgrotte fragments of goethite crusts were found (Fig. 10).

The map (Fig. 9) shows that many of these caves are grouped in the center of the Alb at around Muggendorf on the flanks of the Wiesent valley. The valleys in the northern Franconian Alb are striking mostly NW-SE, a direction called "Hercynian" in Central Europe. Other directions are NNE-SSW ("Rhenish")

Figure 6. View up into one of the pits in the Laichinger Tiefenhöhle showing cupolas, typical for phreatic convective dissolution and indicating that these pits have not been created by vadose seepage water.

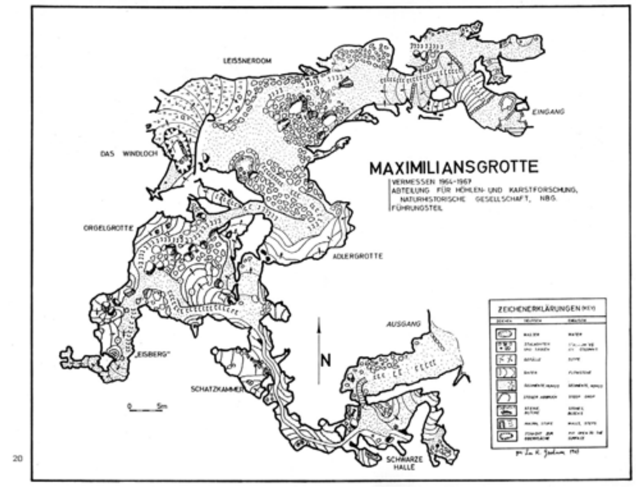

Figure 8. Map of the Maximiliansgrotte characterized by large irregular halls connected by small, irregular passages (Abteilung für Karst- und Höhlenkunde der NHG Nürnberg, 1994).

and NE-SW ("Erzgebirgisch" or "Variscian"). The tectonic domination of the Alb valleys have long been noticed (for an overview see Baier, 2008). Thus, it appears that the site of hypogene caves is controlled tectonically. The Franconian Alb has been compressed and forms a shallow bowl-shaped basin (Streit, 1974, 1977), visible in the direction of the river network that focuses towards the center of the bowl. Thus, it is conceivable that deep-reaching fractures formed that opened migration paths for crustal fluids and gases. Below the Malm, we find a thick package of middle and lower Jurassic pyrite-rich clays. Thus, sulfide-rich fluids could have been mobilized from below causing the observed hypogene karstification. Further down a thick suite of Triassic rocks occurs, including the middle Muschelkalk evaporites that could have functioned as methane trap that were tapped by the fractures. The methane could have evolved from the lower Muschelkalk, a marine marly limestone. Thus, several options exist, to obtain reducing gases. That CO_2 is not an option can be concluded from the fact that the discussed caves are not found at the base of the Malm but in its upper layers. Also they are more horizontal than vertical in shape suggesting control by the oxygen front in the former groundwater body. This scenario is substantiated by the finding of a substantial goethite crust in the Bismarckgrotte by the author (Fig. 10). These may even have had the form of chimneys, such as if they represented "black smokers." Unfortunately, these crusts have been much destroyed by the traffic through the cave. If such crusts appeared at the bot-

Figure 7. Map of the Zoolithenhöhle characterized by irregular halls connected by pits and narrow passages (Heller, 1972).

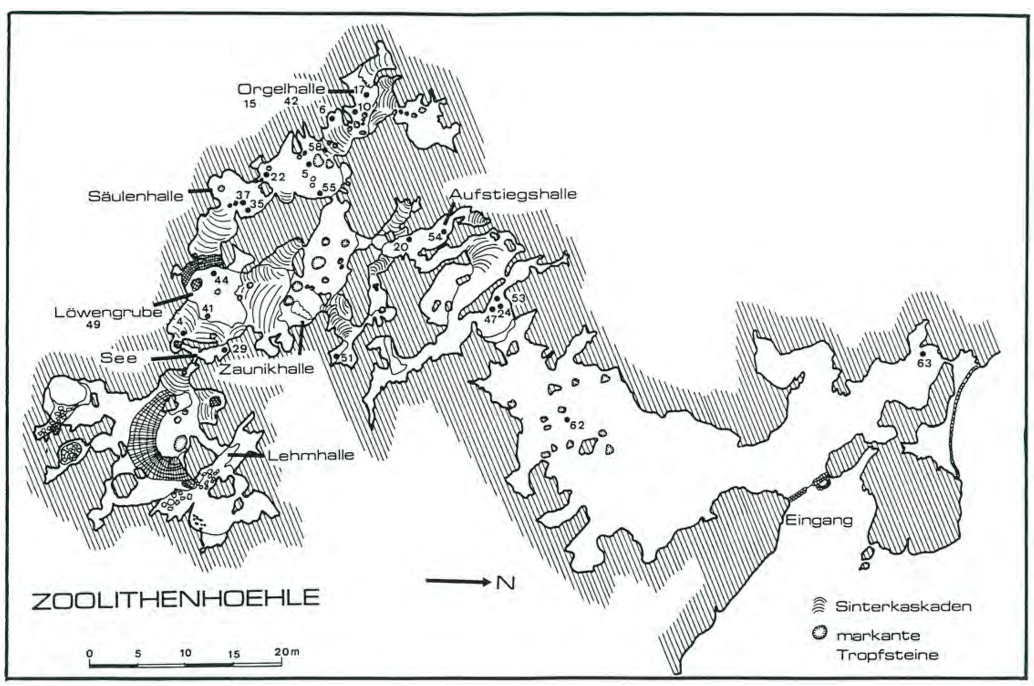

Figure 9. Google Earth view with some of the larger hypogene caves in the Franconian Alb pinned. Blue: Outline of the Malm escarpment; yellow: Hercynian lineaments; green: other lineaments.

Figure 10. Piece of goethite/limonite with a vertical texture from the bottom of the Bismarckgrotte. It could suggest that anaerobic waters (high in ferrous iron solutions) entered the cave to be oxygenated, thereby producing acidic byproducts that corroded the cave and precipitated the goethite in situ.

tom of other caves as well cannot be ascertained. This is because German caves are generally filled by glacial solifluction loams. In case of the Bismarckgrotte, the site where these crusts occur is away from the entrance pits, so that it was not covered by later sediments.

The cave at Stein-am-Wasser (actually called "Höhle ohne Namen"), a large maze cave, may be a different case. Here we measured a CO_2 concentration of up to 1 % in the passages furthest away from the entrance, a value much higher than normally found in our caves (Kempe et al., 1998). Since the water table is intersecting the cave, it appears as if it is the source of this high CO_2-value.

The southern-most karst areas of Germany are represented by the few Alpine carbonate mountain massifs along the border with Austria. These areas seem to have a rather high cave density. Here the longest and deepest caves in Germany (Riesending, 18 km long, 1,059 m deep; Hölloch 11 km long, 452 m deep; VdHK, 2013) have recently been explored. These caves seem to be series of vadose shafts connected by epigene canyons and not so much hypogene.

CONCLUSIONS

This overview over German carbonate rock caves is intended to trigger discussion about the classification of these caves. Most of them have been known for a long time but have never been viewed in context. Here I attempt to show that (i) hypogene caves are very common in Germany and (ii) that some of them form well-defined clusters, such as at the Iberg/Harz, or in the Franconian Alb. Other cave areas seem to show mixed evidence with epigene caves formed side-by-side with caves of tentative hypogene origin, such as in Elbingerode/Harz, in the Rhenish Schist Massif and in the Swabian Alb. Except for the Iberg, where siderite seems to provide the cave-forming agent, the cause of hypogene cave formation remains undecided. It could be anything from rising volcanic CO_2 (as in the Swabian Alb) or any of the other gases, methane or hydrogen sulfide (in case of the gigantic 100 million years old Wülfrath cavity). We yet have to wait for some lucky find (such as gypsum preserved in some dry, obscure corner of a cave) or a method to extract information from initial autochthonous sediments as to the leaching conditions that formed the caves millions of years ago.

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HOW DEEP IS HYPOGENE? GYPSUM CAVES IN THE SOUTH HARZ

***Stephan Kempe*[1]**

Germany currently features 20 caves in sulfate rocks (gypsum and anhydrite) longer than 200 m. Most of them occur either in the Werra-Anhydrite or in the Hauptanhydrite of the evaporitic Zechstein series (Upper Permian). One occurs in the Jurassic Münder Mergel and two in the Triassic Grundgips. The longest, the Wimmelburger Schlotten, is 2.8 km long with a floor area of 24,000 m2. All caves, except four, occur in the South Harz, where the Zechstein outcrop fringes the uplifted and tilted Variscian Harz. These caves can be divided into three general classes: (i) epigenic caves with lateral, turbulent water flow, and (ii) shallow or (iii) deep phreatic caves with slow convective density-driven dissolution. The latter were discovered during historic copper-shale mining and called "Schlotten" by the miners; most of them are not accessible any more. Shallow phreatic caves occur in several areas, most notably in the Nature Preserve of the Hainholz/Beierstein at Düna/Osterode/Lower Saxony. Here, we sampled all water bodies in May 1973 and monitored 31 stations between Nov. 23rd, 1974, and April 24th, 1976, with a total 933 samples, allowing us to characterize the provenance of these waters. These monitoring results were published only partially (PCO_2 data, see Kempe, 1992). Here, I use the data set to show that the Jettenhöhle (the largest cave in the Hainholz) has been created by upward moving, carbonate-bearing, groundwater of high PCO_2. Even though the cave has now only small cave ponds and essentially is a dry cave above the ground water level, it is a hypogene cave because of the upward movement "of the cave-forming agent" (sensu Klimchouk, 2012). Likewise, the Schlotten are created by water rising from the underlying carbonate aquifer, but under a deep phreatic setting.

THE QUESTION

Hypogene karstification can be defined in several ways (for a recent overview see Klimchouk, 2012). Central for it is that "the cave-forming agent" (Klimchouk, 2012, p.750) responsible for cave formation rises from below. In case of hypogene caves in evaporative rocks, the "agent" is the water itself. It rises from strata below the karstifiable rock and either returns there or emerges as artesian water to the surface dissolving the evaporites (gypsum, anhydrite or salt) as it passes through.

All this normally should occur below or even far below the local water table in the phreatic realm. But what if the majority of the cave volume formed is largely above the water table even though it is not fossil? This is the case for some of the German gypsum caves.

GYPSUM CAVES OF GERMANY

Gypsum or anhydrite caves are relatively rare. Only a few countries feature sulfate caves in any appreciable number, among them are Canada, the USA, Spain, France, Italy, Switzerland, Austria, Germany, the Ukraine, Moldova, Russia, Syria, and Saudi Arabia. Among sulfate caves, the five longest ones occur in the Ukraine and Moldova, with the maze cave Optymistychna leading the list with 236 km in length (see list of longest gypsum caves, edited by Gulden, 2013).

The gypsum caves of Germany are among those attracting the scientific curiosity first (e.g., Behrens, 1703). Even though more than half of the German area is underlain by the evaporites of the

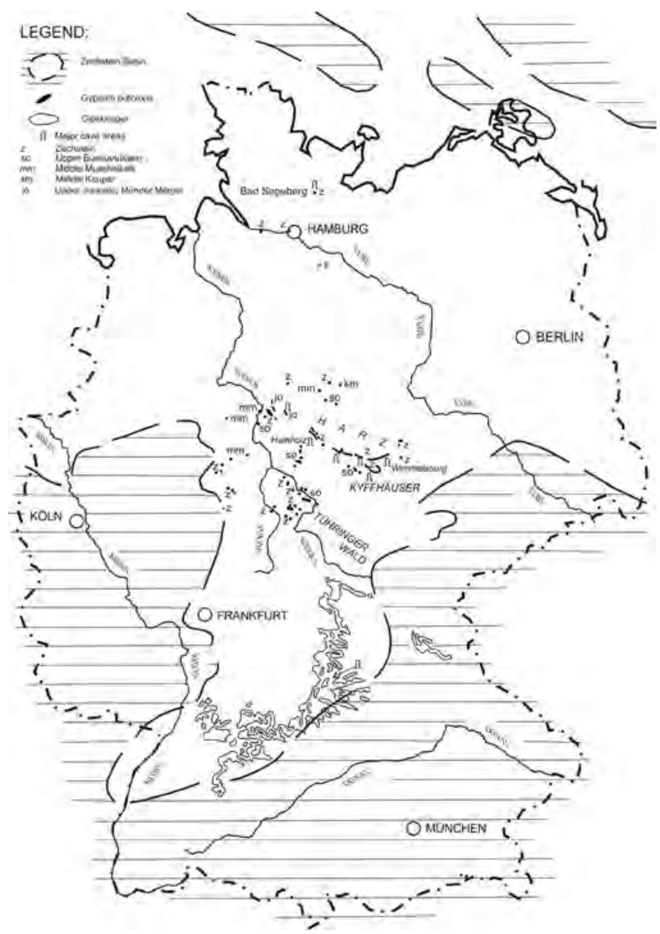

Figure 1. Zechstein Basin (white area) underlaying Germany in large parts. The second largest area is occupied by the Keuper (Triassic) gypsum layer in Southern Germany (fragmented area to the east and south of Frankfurt) (Kempe, 1996).

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Zechstein (today a litho-stratigraphic group roughly equivalent of the Upper Permian; equivalent to the Lopoginium Formation, 260 to 252 Mio year ago) (Fig. 1), the areas of sulfate outcrops are relatively small (small black lines and dots in Fig. 1). The Zechstein, historically one of the first formations named as such, is, at the center of its basin, composed of up to seven evaporitic cycles. Salt never appears at the surface and only the three oldest sulfate layers (the Werra-Anhydrite, the Sangerhäuser/Basal Anhydrite and the Hauptanhydrite, belonging to the Werra-, Staßfurt-, and Leine-Cycles) accompanied by intercalated limestone, dolomite, and claystone layers form appreciable outcrops. These anhydrite layers are gypsified near the surface. The Werra-Anhydrite can be up to 250 m thick at places.

The largest of the Zechstein-outcrops occurs along the southern rim of the Variscian (Hercynian) folded block of the Harz and Kyffhäuser Mountains (Fig. 1). Here, over 90% of German sulfate caves are located. Table 1 lists the longest sulfate caves in Germany (modified after Kempe and Helbing, 2000) and Fig. 2 gives the distribution of lengths and areas of these caves showing close correlation with third order exponential fits. The areas given in the table show that most of the caves are very large cavities, indeed. Currently, we have laser-scanned two of these caves (Barbarossa Höhle and Himmelreichhöhle) that will eventually allow also the calculation of rather exact volumes.

Table 1 already shows that we can easily differentiate between three types of caves: epigenic (creek-fed, vadose stream caves) and shallow phreatic and deep phreatic caves, both essentially hypogenic in origin.

Table 1 also suggests that all of the deep phreatic sulfate caves were discovered by mining for the underlying, metalliferous copper-shale (a thin, marine, marly claystone at the base of the Werra Cycle containing up to 5% Cu). The copper-shale miners called these cavities "Schlotten," of which we only have a few maps published (Stolberg, 1943). According to Rainer and Christel Völker, who intensively studied "thousands of pages of old mining documents", there may have been hundreds of these caves known but only for a few of them maps are available (Table 1). The caves proved to be a bonus for the miners who could use them to dump mine refuse and to drain water into them. Accordingly, their ownership was worthwhile to fight for in court, thereby leaving documentation. By building deeper and deeper water solving mine passages more and more of these caves appeared. Some are still accessible through mine shafts and with special permits, such as the Wimmelburger Schlotten (Fig. 3), here listed with 2.8 km of passage. This is the length to which the cavities were mapped but much more was apparently accessibly prior to the shut-down of the copper-shale extraction in the Mansfelder Mulde in the 1980's (pers. comm. R. & C. Völker).

Alreadym Johann Carl Freiesleben (1809) understood that the Schlotten formed by convection below the water table. He is thus one of the earliest geologists describing "hypogene" speleogenesis. The general tectonic situation of the Schlotten is given

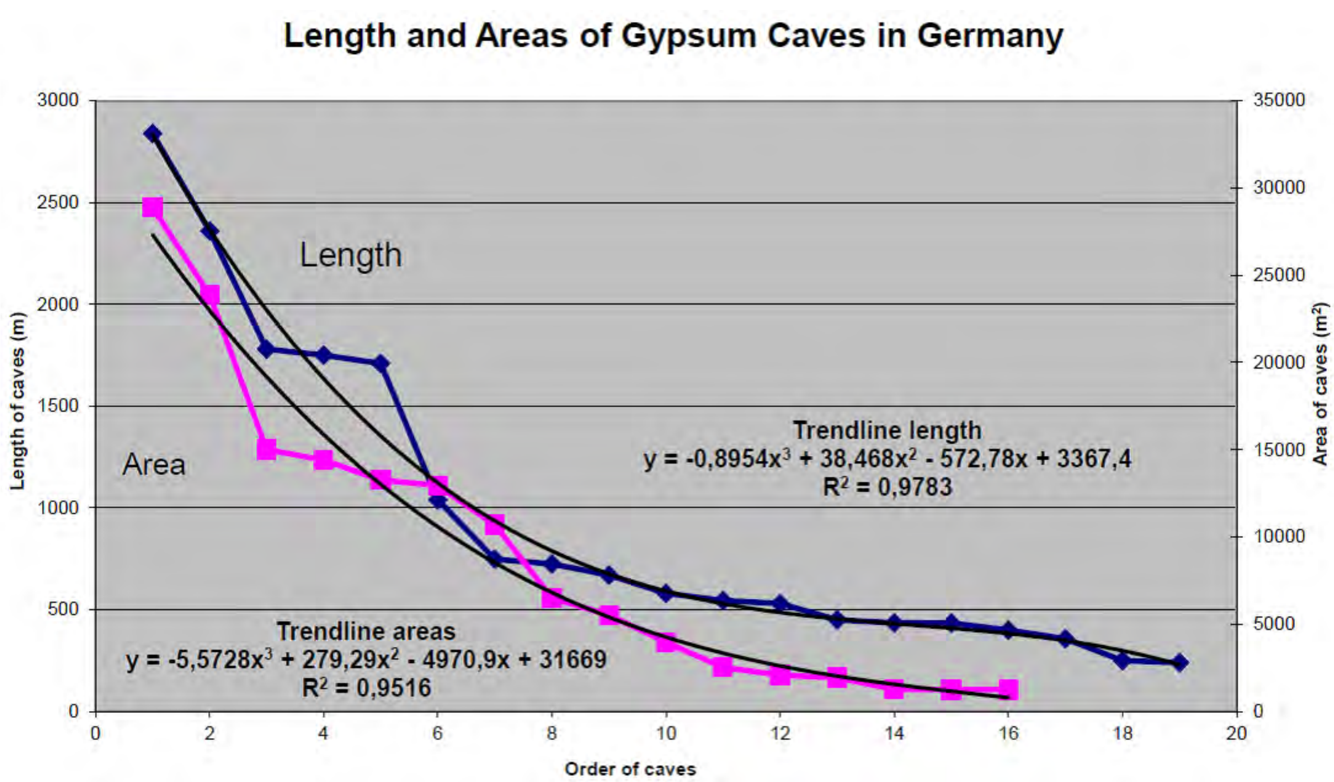

Figure 2. Distribution of lengths and areas German gypsum caves according to Table 1.

Table 1. List of German sulfate caves with more than 200 m length, sorted by length (citations see Kempe, 1996, 1998; data from Kempe and Helbing, 2000).

No.	Name, location and speleogenetic realm	Length, m	Area, m^2
1	Wimmelburger Schlotten**/anhydrite, Wimmelburg, E-Harz, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave system; drained by copper-shale mining (Biese, 1931; Stolberg, 1943; Völker & Völker, 1986)	2,838	23,866
2	Segeberger Kalkberghöhle*/anhydrite & gypsum, Bad Segeberg, Schleswig-Holstein; large, shallow phreatic, hypogene or marginally epigene solution maze cave with some breakdown rooms, drained naturally or anthropogenicly several centuries ago (Gripp, 1913; newly surveyed including all side passages, Fricke,1989)	2,360	6,511
3	Heimkehle*/gypsum, Uftrungen, S-Harz, Saxony-Anhalt; Large, shallow phreatic hypogene and marginally epigene cave dominated by very large breakdown halls, cave still active (Stolberg, 1926; Biese, 1931; Völker, 1981)	1,780	14,407
4	Numburghöhle***/gypsum, Kelbra, Kyffhäuser, Saxony-Anhalt; Very large, shallow phreatic, hypogene solutional cave with very large breakdown halls, cave shortly drained (1989) by coppershale mining, now flooded again (Stolberg, 1926; Völker, 1989; Völker & Völker, 1991)	1,750	28,866
5	Schlotte am Ottiliaeschacht****/anhydrite? Ahlsdorf, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave, drained by copper-shale mining, likely flooded today (Stolberg, 1943)	1,710	15,009
6	Höllern/gypsum**(Keuper), Markt Nordheim, Franconia, Bavaria; Low, shallow phreatic, hypogene solutional maze cave composed of low and narrow passages, active (Cramer & Heller, 1933; Götz, 1979)	1,040	2,531
7	Jettenhöhle-Kleine Jettenhöhle/gypsum, Hainholz, S-Harz, Lower-Saxony; Large, shallow phreatic, hypogene solutional cave, dominated by large breakdown halls, active (Stolberg, 1926; Kempe et al., 1972, enlarged by 130 m since 1990)	748	n.d.
8	Schlotte am Schacht E**** /anhydrite? Ahlsdorf, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave, drained by copper-shale mining, likely flooded today (Stolberg, 1943)	725	5,525
9	Barbarossahöhle*/anhydrite, Rottleben, Kyffhäuser, Thuringia; Large, shallow phreatic solutional cave, dominated by large breakdown halls (Biese, 1923; Kupetz & Mucke, 1989; Kupetz & Brust, eds., 1994)	670	12,946
10	Himmelreichhöhle**/anhydrite, Walkenried-Ellrich, S-Harz, Lower-Saxony; Large, epigene cave, domiated by a very large breakdown hall passed by a railway tunnel and drained in the 19th century by water tunnels (Biese, 1931; Reinboth, 1970)	580	10,695
11	Fitzmühlen Quellhöhle/gypsum, Tettenborn, S-Harz, Lower-Saxony; Low, epigene active cave (Haase, 1936; map by Hartwig, 1988, unpublished)	545	1,253
12	Brandschächter Schlotte****anhydrite? Sangerhausen, S-Harz, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave, drained by copper-shale mining, likely flooded today (Stolberg, 1943; Völker, R., 1983)	530	13,267
13	Marthahöhle**/gypsum, Hainholz, S-Harz, Lower-Saxony; Shallow phreatic, hypogene, marginally epigene solutional cave, active (Stolberg, 1936; Kempe et al., 1972)	450	3,969
14	Großes Trogstein System***/gypsum , Tettenborn, S-Harz, Lower-Saxony; Mostly low epigene cave, feeds into Fitzmühlen Quellhöhle (Stolberg, 1928, 1932; Biese, 1931; Reinboth, 1963, 1969)	435	1,945
15	Schusterhöhle**/gypsum, Tilleda, Kyffhäuser, Saxony-Anhalt; Shallow phreatic solutional cave (R. & C. Völker, pers. comm.)	434	n.d.
16	Schlotte am Eduardschacht****/anhydrite ? Mansfeld, E-Harz, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave, drained by copper-shale mining, likely flooded today (Kupetz & Brust, 1991)	ca.400	n.d.
17	Elisabethschächter Schlotte**/anhydrite, Sangerhausen, S-Harz, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave, drained by copper-shale mining (Stolberg, 1943; Völker & Völker, 1982a)	357	2,082
18	Mathildenhöhle/gypsum (Jurassic, Münder Mergel), Ammensen, Holzminden, Lower-Saxony; Active epigene cave (pers. comm. S. Meyer)	317	1,250
19	Höhle im Grundgips der Kläranlage****/gypsum (Keuper), Bad Windsheim, Franconia, Bavaria; Probably epigene cave	250	n.d.
20	Segen Gottes Schlotte**/anhydrite, Sangerhausen, S-Harz, Saxony-Anhalt; Large, deep phreatic, hypogene solutional cave, drained by copper-shale mining (Stolberg, 1943; Völker & Völker, 1982b)	240	1,279

* = show-cave; ** = accessible with permission (essentially inaccessible); *** = major part inaccessible; **** = not accessible any more (flooded); n.d. = not determined

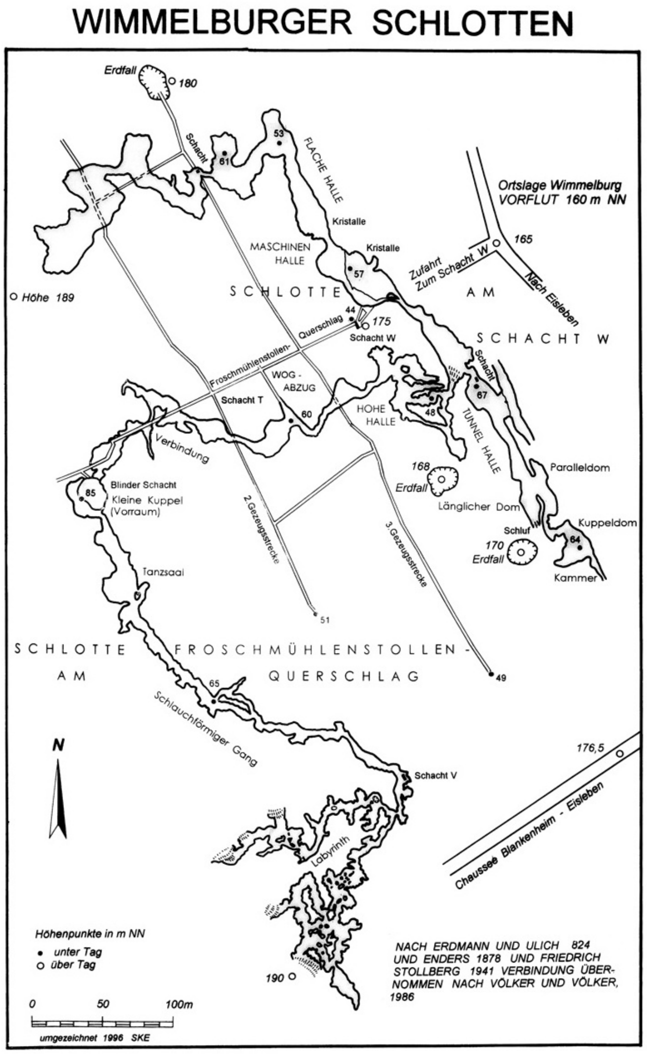

Figure 3. Map of the Wimmelburger Schlotten, the largest deep-phreatic cave system, originally discovered and drained by mining for the underlying copper shale and partly still accessible by (rarely granted) permit (Kempe, 1996).

in Figure 4d, showing that the “dissolving agent,” i.e. calcium sulfate undersaturated water, derives from water sinking in the underlying Zechsteinkalk, a 6 to 10 m thick package of well-jointed micritic limestone beds. Because of its lower buoyancy, the water will rise into the hanging Werra Anhydrite, dissolving it and thereby becoming denser, and consequently returning into the limestone layer driving up additional, less-saturated water. Thus, by and by large convective cavities can be formed at the base of the Werra Anhydrite.

The geology of the Zechstein-girdle along the southern rim of the Harz is currently the subject of a doctoral thesis by Hans-Peter Hubrich. He is compiling all geological maps, including detailed maps from over 40 diploma theses and more than 20 years of student mapping courses under the direction of the author. These maps show that the Zechstein is tectonically highly fractured (Fig. 4b, c). The general dip of the strata of 10-15° to the SE (caused by the uplift of the Harz block) should allow only for a relatively narrow outcrop forming a classical set of escarpments (Fig. 4a). In most areas, however, multiple NW-SE striking faults cause repeated outcrops of the same strata, thereby broadening the gypsum karst area considerably to several kilometers (Fig. 4c). These settings allow for a wide variety of hydrological settings for karstification.

THE CAVES OF THE HAINHOLZ/ BEIERSTEIN AT DÜNA/OSTERODE

None of the Schlotten are available for detailed hydrological or geochemical studies. Thus the explanation for their genesis suggested above and in Fig. 4d remains hypothetical at this time. However, the Hainholz at Düna near Osterode (Fig. 5), since 1970 a nature reserve (Kempe, 1972), offers a possibility to study the involved processes. Here, the Hauptanhydrite is situated in a graben position that protected it from rapid erosion (Fig. 4b).

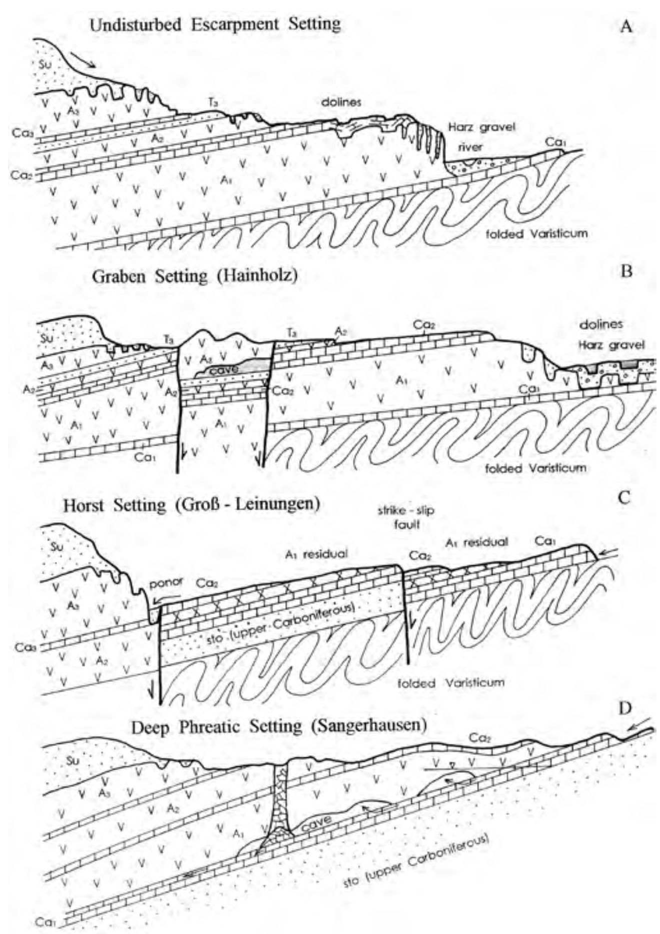

Figure 4. Various tectonic situations of the South Harz, fostering different speleogenetic setting (Kempe, 1996).

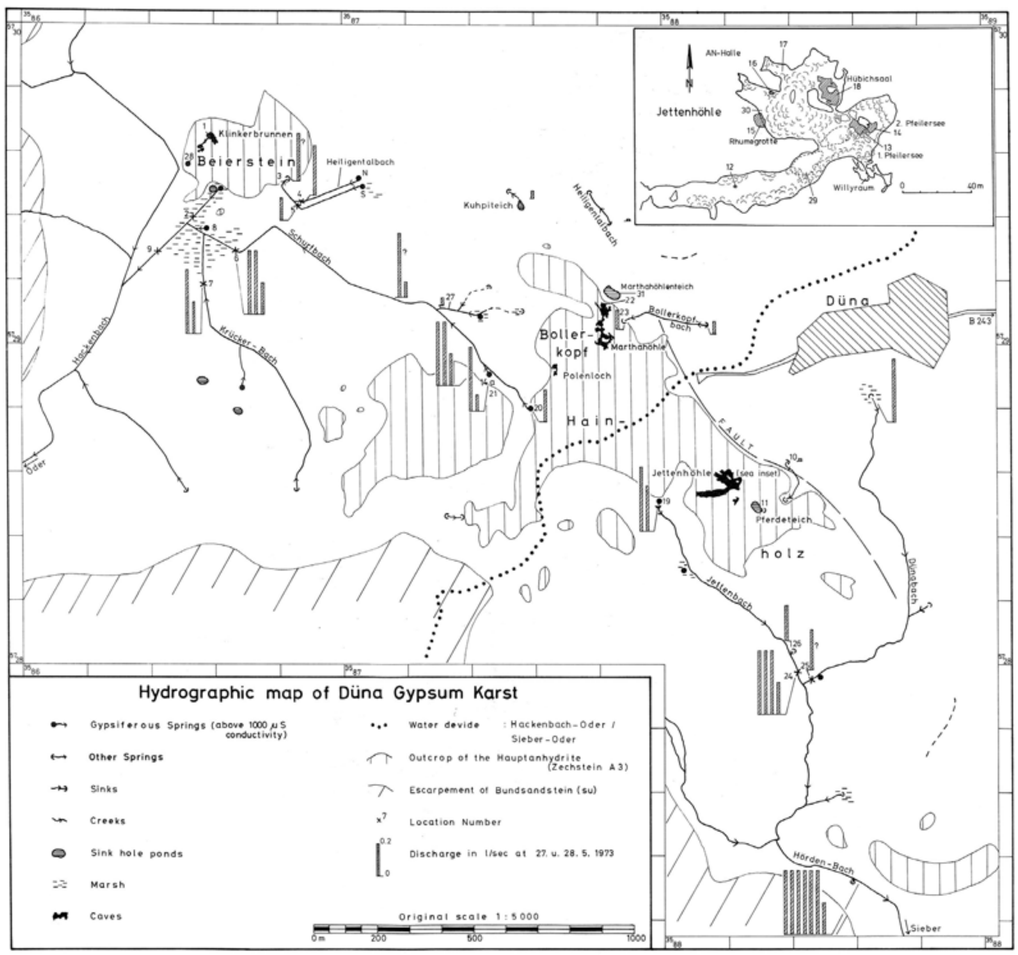

Figure 5. Map of the Düna gypsum karst area (Hainholz and Beierstein), its main fault, its hydrographic features and the sampling points (Kempe, 1992). Inset shows the location of the sampling points within the Jettenhöhle.

The preserve features more than 20 gypsum caves, several of them are still active shallow phreatic solutional caves. The major ones are the Jettenhöhle/Kleine Jettenhöhle (Table 1, Fig. 6), the Marthahöhle (Table 1) (Fig. 7), the Klinkerbrunnen and the Polenloch. Other caves are associated with sinkholes, sinking creeks, or are fractures enlarged by seepage water (Kempe et al., 1972).

In May 1973, the Arbeitsgemeinschaft niedersächsische Höhlen (ArgeNH) sampled all available cave ponds, sinkhole ponds, springs and creeks in the area (Fig. 5) (Brandt et al., 1975). This sampling provided data that allowed classifying the various karst water bodies. In continuation of this work, the ArgeNH sampled most of these sites on a biweekly basis (i.e., 35 series) between Nov. 23rd, 1974, and April 24th, 1976. Hydrochemical results of 933 samples of 31 stations were published by Kempe (1992) as part of his habilitation thesis. Data at the time were mostly evaluated for PCO_2 (CO_2 pressure), not so much for total ions and mineral saturation. Methods are discussed in Brandt et al. (1975) and Kempe (1992), and there are also lists with all the results can be found.

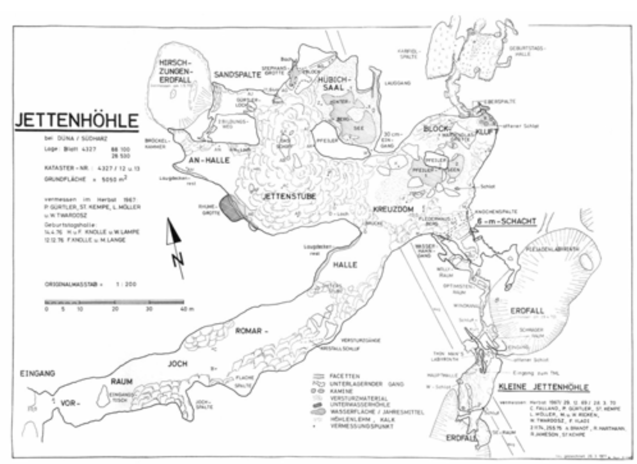

Figure 6. The Jettenhöhle/Kleine Jettenhöhle, the largest cave in the Hainholz area (not all passages plotted). The cave system features only a found groundwater outcrops and is a state of late development (Kempe, 1996) (map by ArgeNH).

Figure 8 plots sulfate versus alkalinity values that allows grouping the waters according to their provenience. Lowest sulfate values but highest alkalinities are characteristic for springs from carbonate karst. Spring 10 (called the "Arteser", since it rises from a 2 m deep natural well) has the lowest Saturation Index of gypsum (SIgyp) of -1.825 ± 0.592, N = 19, and the highest PCO_2 (20,015 ± 9926ppm, N = 30). Its water rises from the Staßfurt limestone (a formation called "Stinkschiefer" because of the intense oily smell when broken) east of the Hainholz Fault that down-thrusts the Hauptanhydrite to the west (see Fig. 5). The spring of the "Bollerkopf Creek" delivers water from the same strata and has similarly low sulfate (SIgyp -1.615 ± 0.471, N=19) and high alkalinity values but because the collection site was several hundred meters below the spring, the PCO_2 is much lower (2512 ± 1807ppm N = 28) than in the "Arteser". Both samples have very low total dissolved solids (TDS) (519 ± 151 and 511 ± 135 mg/l, respectively).

Highest sulfate and lowest alkalinity values are displayed by seepage water collected in the Jettenhöhle (No 12), with a mean SIgyp of -0.024 ± 0.046, N=11, and a TDS of 2120 ± 180 mg/l. Associated shallow cave ponds St. 29 and 30 have SIgyp of -0.082 ± 0.079, N = 7, and -0.105 ± 0.095, N = 7, and TDS values of 1863 ± 478, N = 21, and 2039 ± 366 mg/l, N = 18, respectively. Interestingly, the seepage water is very low in PCO_2, i.e. 735 ± 497ppm, N = 21, showing that the water degassed largely during dripping, while the seepage water pools show higher PCO2 (St. 29 1347 ± 1133ppm, N 18; St. 30 2727 ± 1719ppm, N = 14) suggesting that these puddles are fed also by water films and not only by freely dripping water.

In between these end-member groups of waters dominated by alkalinity and sulfate, we find two other groups, i.e. those samples that represent the sink-hole ponds that are characterized by relatively high alkalinities and large variations in concentrations, and those that represent cave ponds and sulfate-karst springs which have lower alkalinities and higher sulfate concentrations and lower variations in concentrations (Fig. 8).

Figure 9 shows the interannual TDS variations of some of the stations in comparison. As already stated, the "Arteser" (St. 10) has the lowest values throughout the sampling period. Moreover it does not display any interannual trend. The highest values are displayed by the seepage water of sample 12. Its record was interrupted between end of August and end of December 1975 because of dry weather conditions. Afterwards, the TDS values were lower, possibly because of rapid winter rains, percolating fast through the cracks of the cave's ceiling. The second highest values belong to the "Jettenquelle," a small spring (St. 19), opposite the entrance to the Jettenhöhle but most probably hydrologically not connected to it. Its average TDS amounted to 2020 ± 428 mg/l, N 34, and its SI_{gyp} to -0.092 ± 0.096, N = 18. Below these two highly saturated stations two of the prominent cave ponds of the Jettenhöhle are plotted, the 1. Pfeilersee (St. 13; a less than 2 m deep pool at the pillar separating the "Kreuzdom" from the "Blockkluft"; see Fig. 6) and the Hübichsaal pond (St.

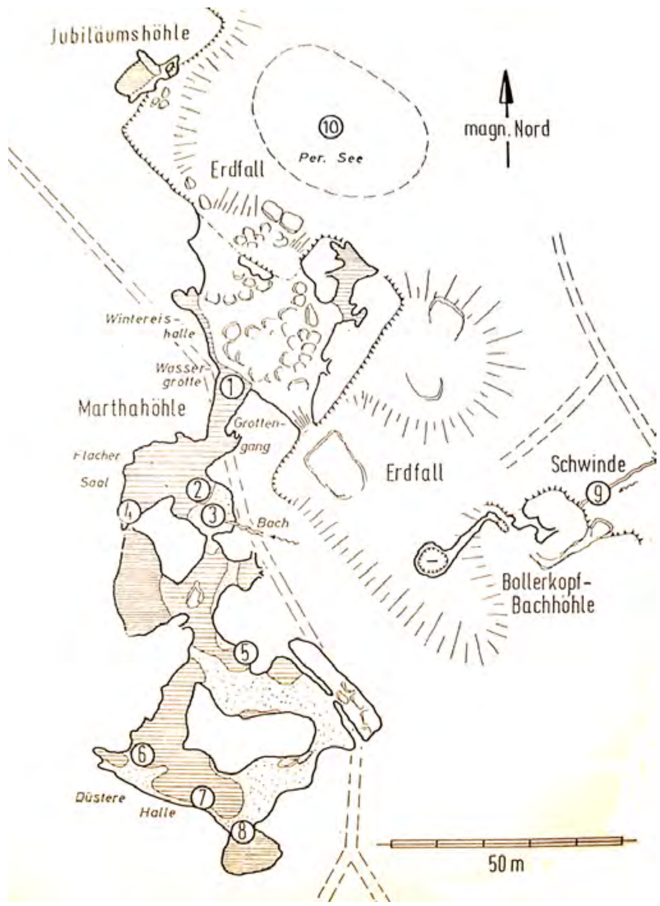

Figure 7. The Marthahöhle, the second largest caves of the Hainholz. Apart from the breakdown-dominated entrance hall, the cave is accessible only during very low ground water levels (Map redrawn after Stolberg, 1936).

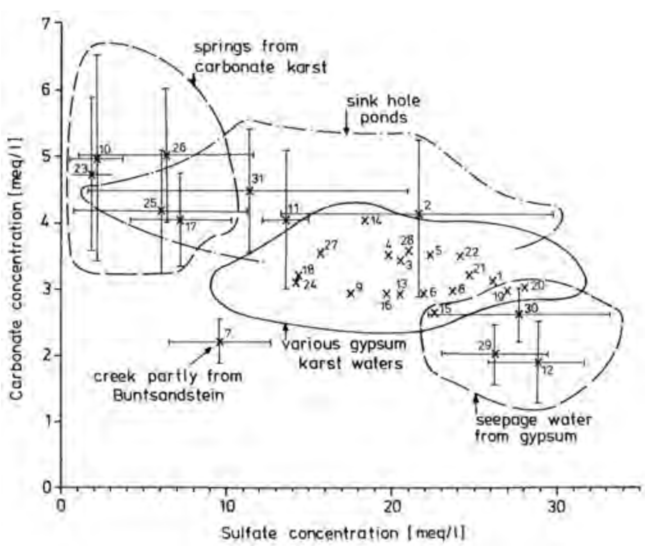

Figure 8. Average sulfate versus carbonate (alkalinity) concentrations of the long-term sampling of 31 locations of the Hainholz and Beierstein. In the gypsum karst field, the standard deviations are omitted for greater clarity.

18; a less than 1.5 m deep pond in the Hübichsaal; see Fig. 6) that have averaged TDS concentrations of 1629 ±3 53mg/l, N = 34, and 1201 ± 368 mg /l, N = 35, and SI_{gyp} values of -0.261 ± 0.153, N = 21, and -0.489 ± 0.199, N = 21, respectively. These values are much less than those of the seepage water, suggesting that these ponds have collected not seepage waters but upwelling groundwater. To substantiate this conclusion, the PCO_2 values are much higher than that of the seepage water, i.e. 3019 ± 2266ppm, N = 30, and 4029 ± 2026ppm, N = 1, again suggesting upwelling from below. The concentration relations are similar also for the other cave ponds of the Jettenhöhle: The 2. Pfeilersee (St. 14) and the Rhumegrotte (St. 15) and the internal spring in the AN-Halle (St. 16.) all have lower TDS and SI_{gyp} values and higher PCO_2 values than seepage water.

Interesting is also that the concentrations of the four upper curves decreased towards the winter of 1975. The ceasing of the seepage water suggests a dry period. Apparently, then the cave ponds received less seepage water causing a lowering in TDS because they contain a higher proportion of upwelling groundwater.

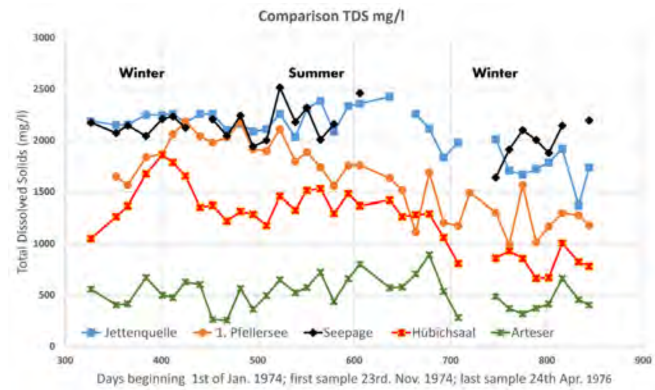

Figure 9. Interannual variation of TDS of some stations of the Hainholz and Beierstein sampling program.

CONCLUSIONS

The investigation of the Hainholz illustrates that even in caves that appear to be vadose, i.e. having most or much of its volume above the water table, the “cave-forming agent” can come from below. In the Zechstein belt, that accompanies the tilted Variscian Harz massif along its south, three anhydrite layers crop out, each underlain by limestone beds. These can function as aquifers, in which water of high alkalinity and high PCO_2 can circulate. According to the measurements of the Hainholz/Beierstein monitoring from 1974 to 1976 the TDS of these waters reaches concentrations of 500 to 600 mg/L. In contrast to this, water that seeps through the gypsum of the roof of the Jettenhöhle that is between less than 10 m and 15 m thick reaches a concentration of above 2000 mg/L, very near to saturation of gypsum, and has a low PCO_2. Thus seepage water cannot be “the cave-forming agent”. Rather ascending groundwater, present in the various cave ponds and originating in the underlying limestone layer (the Plattendolomit of the Leine Series in case of the Jettenhöhle) and strongly undersaturated with respect to gypsum, is responsible for the cave genesis. The Jettenhöhle is apparently in its late stage, enlarged by breakdown that fell into the groundwater body and became dissolved. In this manner the cave ceiling grew well above the level of the karst water table. At the same time residuals and, probably even more important, calcite that was forced out of solution by the ongoing gypsum dissolution, collected on the floor. Drilling showed that these sediments are more than 2 m thick and contain layers of calcite rafts, typical for degassing cave ponds. Thus the cave today is dry throughout most parts. This is in contrast to the Marthahöhle and the Klinkerbrunnen that are accessible only during periods of prolonged drought. Otherwise, the caves are water-filled and display the typical morphological features of caves developing by natural convection: flat solution ceilings (Laugdecken) with solution cups (Laugnäpfe) and sloping sidewalls (Facetten) (e.g., Kempe, 1996). Also, unlike the Jettenhöhle that has no input by a sinking creek nearby, the Marthahöhle receives water (when active) from the Bollerkopf creek while the Beierstein receives water through the Beierstein ponor. Nevertheless, these caves are by no means epigene because of their morphology. The Polenloch is in an intermediate state: there a large solution cavity has collapsed and the breakdown mountain is still in place, fringed by groundwater pools.

In case of the deep-phreatic Schlotten, discovered by the copper-shale mining, the groundwater percolated through the Zechsteinkalk, to dissolve the Werra-Anhydrite from below. The gypsum-saturated water then returned into the Zechsteinkalk in order to disappear into deeper groundwater levels where rock salt Zechstein series awaited their turn to be dissolved at depths of around 200 m below the surface.

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Over the years many friends and colleagues cooperated in the exploration of the South Harz gypsum karst. Specifically Firouz Vladi, Rainer Hartmann, Martin Seeger, Andreas Brandt, and Eric Barran have to be acknowledged, among others, in conducting the regular sampling of the Hainholz-Beierstein. This paper is dedicated to them.

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RESERVOIR CHARACTERISTICS OF THE COMPLEX KARST OF THE LLUCMAJOR PLATFORM, MALLORCA ISLAND (SPAIN): TOOL FOR HYDROCARBON RESERVOIR APPRAISAL

Philippe A. Lapointe[1]

The development of porosity in carbonate platforms takes many forms. Dissolution porosity as a result of karst processes is unique as it produces organized porosity and permeability over a variety of scales, and can do so in very short periods of time, geologically speaking. Karst developed in the Miocene formations of the Mallorca Island exhibits a complexity that seems to be very similar to the Kashagan or Aktote (Kazakhstan) or Kharyaga (CIS) karst reservoirs architecture characterized by different phases of island karst (mixing water) type with caves of different sizes and sponge karst, reworked and partly filled by paleosoils related to plateau karst developed during major sea level drops and finally hydro- (geo)-thermal processes. The Miocene rocks of the Llucmajor platform in the southwest of Mallorca island exhibit the three main types of karst developments that occurred through time, linked or not to glacio-eustatic changes: -1 Island karst (the flank-margin model); -2 Meteoric karst; -3 Hydrothermal karst/ These developments allow defining the so-called Complex Karst. Each of the terms is identified by specific overprints found in drilled wells (logs and cores) or on outcrops. The outcrops and subcrops of Mallorca Island represent an excellent analogue for understanding the complexity of the past carbonate platforms which are hydrocarbon targets for the industry.

LOCATION

Reef-rimmed prograding platforms were widespread in the western Mediterranean during the Late Miocene (Esteban, 1979). The study area is located south-east of Mallorca, on the Late Miocene Llucmajor platform, which is 20 km wide and up to 200 m thick (Esteban, 1979; Pomar et al., 1983) (Fig. 1). The Llucmajor platform was formed from the Late Burdigalian to Early Messinian. The depositional sequences are characterized by a shallowing-upward trend, with alternating phases of progradation and aggradation related to sea-level fluctuations.

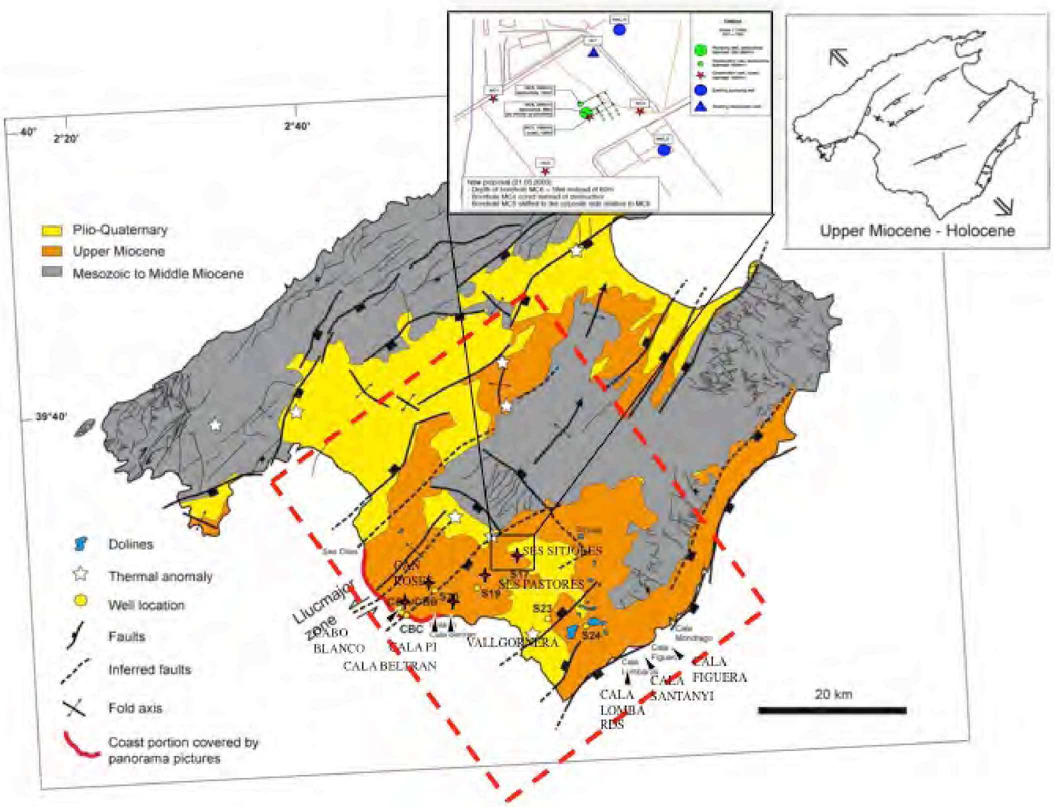

Figure 1. General geological map and location of study area

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GEOLOGY OVERVIEW

The Upper Miocene layers are fairly flat-lying limestones and dolostones, and have undergone only slight tilting and flexure associated with normal and strike-slip faulting during Late Neogene to Middle Pleistocene times. Our approach to the Miocene carbonate systems is based on the study of outcrops and new drill cores that have been carried out in the framework of two programs: the "ISES-Shell" Program (J. Kenter coord. unpublished) and the "ALIANCE" Program (Ph. Pézard coord. 2007). The rocks are composed of three major sedimentary units that correspond to third-order depositional sequences (Pomar and Ward, 1994; Pomar et al., 1996; Pomar et al., 2004) with:

- Lower sequence (Early Tortonian): carbonate ramp, including Heterostegina calcisiltites and rhodalgal lithofacies without coral reefs;
- Middle sequence (Late Tortonian – Early Messinian): well-developed progradational reefal platforms;
- Upper sequence (Messinian): variety of lithologies including oolites and stromatolites.

HYDROLOGY

The Llucmajor platform contains a fresh water reservoir affected by an extensive seawater intrusion (Maria-Sube, 2008). At the Campos site, located ~7 km inland, about 20 m of brackish water overlies the salt water. In between, a ~17-m-thick diffusion zone is observed (mixing of fresh and salt water). Four hydrological zones can be distinguished from top to bottom: 1) vadose zone with meteoric water percolation, 2) brackish zone, 3) transition zone and 4) sea water zone (Hebert, 2011). Hydrothermal water corresponding to a deep seated aquifer recharged by fresh water infiltration on the western side of Mallorca flows to the surface through thermal spring (Garcia et al., 2012).

KARST SYSTEM

The limestone consists of a mixture of aragonite, high- and low-magnesium calcite, pure calcite and dolomite, minerals highly reactive on the fluid composition, the calcite and magnesium concentration, the salinity, and the pH. According to the type of fluid-rock interactions, post-deposition diagenesis includes: cementation, dissolution, dolomitization, compaction and recrystallization (Moore, 1989). The overall result is a complex karst system. The main objective was to study a complex karst system analog for application to Kaszakhstan and other karstic petroleum reservoirs. The karst developed in the Miocene formations of Mallorca Island exhibits a complexity looking very similar to the Kashagan and Aktote Karst reservoir complexity with different phases of island karst (mixing water) type with caves of different sizes and sponge karst, reworked and partly filled by plateau karst developed during major sea level drops and finally hydro (geo) thermal karst.

Database

Outcrops - The rather flat and low topography in southern Mallorca implies that the Miocene carbonate platform deposits crop out along cliffs on the coastline. These coastal outcrops include Badia Blava, Cabo Blanco, (Cap de Paret d'Es) Puig des Ros, Cala Beltran, Cala Blava, Cala Figuera, Cala Lombards, Cala Pi, Cala Santanyi, Cala Vallgornera, Portals Vells, and Porto Pi.

Drill cores - The southern part of the Llucmajor Platform (Fig.1) was drilled at Cabo Blanco and in the Can Roses area (S20 core). The Campos Basin was drilled in several areas, Ses Sitjoles (MC2 and MC10 cores), Ses Pastores (S19 core). The distance between the drilling zones ranges from 4 to 10 km. In the Ses Sitjoles area, the Miocene platform carbonate forms a fresh water reservoir. Twelve 100 m holes were drilled/cored at the Ses Sitjoles site (a 100 m scale square field), located 6 km off the Mediterranean coast and 20 km away from the outcrops and drilled holes of the Cabo Blanco (CBB and CBC cores).

KARSTIC FEATURES

Field work was conducted on Mallorca to characterize the position, dimension, and distribution of megapores in the coastal cliffs of the southern coast. The position of the voids within the stratigraphic framework was established, and the nature of regular porosity, and spongework porosity, between the megapores was documented.

Observation of inaccessible cliffs of the Southern Mallorca coast, accomplished by boat, shows classic Sea Caves at the base of cliffs, tafoni as well as Flank margin caves opened by cliff retreat.

Sea cavse and tafoni are erosion related and do not belong to the karst system. tafoni geomorphology designates a hollow shape rounded several decimeters to several meters, carved by erosion in the coastal carbonates; sea caves, also known as littoral caves, are a type of cave formed primarily by wave action. A sea cave is formed by erosion of a weak zone along e.g. fracture/fissure/parting, etc., in the coastal carbonates.

Observation of small cliffs accessible on the coast or through a cala (term for a small bay or creek in Mediterranean islands) shows typical flank margin caves (*sensu* Mylroie et al., 1990) and sponge karst (*sensu* Baceta et al., 2002, 2007, 2008), collapsed caves (Robledo et al., 2002, Ardila, 2005). Meteoric small caves or meteoric reworking of flank margin caves are characterized by speleothems.

On the outcrops of the Cabo Blanco, Cala Pi and Cala Figuera, a comprehensive project was devised with the quantitative description and mapping of the karst systems performed by J. Mylroie and his team from Mississippi State University. All cave and karst features were individually surveyed by compass and tape, and then were tied into a survey line backed up by GPS waypoints.

At Cala Figuera and Cala Pi, the flank margin caves were identified with simple chambers, to clusters of chambers. Marine flooding of the Cala creates flank margin caves in the Cala walls. Following sea-level fall and erosional retreat of the coastline, Cala walls expose the flank margin caves and their meteoric speleothems. Vertical connection through fractures, enlarged by late meteoric karst, crosscut and rework flank margin caves as well as potential sponge karst. At least two main levels of flank margin caves were identified. Collapses are frequently seen on the coastline of the studied area. They develop between two specific levels the basis corresponding to the first flank margin caves level and the top corresponding to the top Miocene.

At Cala Figuera, the sponge karst exists at different locations, on both sides of the Cala. Its thickness ranges from 1 to 2 m maximum, with a pretty wide distribution of several 100 m^2. This sponge karst develops as a zone with levels of discrete spongy porosity traceable for several 10 to 100 m inboard. It is located around the mixing zone and possibly linked to microbially mediated sulphide activity in a zone of reduced circulation. It has been reworked by recent meteoric karst as evidenced by the presence of recent speleothems and paleosoil fills. Paleosoil consists of clays and silts filling in most of the spongy porosity.

Four drill cores, namely; MC-2, MC-10, S-19, and S-20, were analyzed in detail for sedimentology and karst characterization with emphasis on overall porosity. Dating on oyster shells was performed for comparison with the stratigraphic canvas and assessing the drill cores stratigraphic position with respect to the coastal cross-section and the flank margin caves location. Cavernous levels were identified and related to the present and paleo aquifer level as well as to a mixing water wedge. This methodology allows defining the porosity of the karst system away from major caverns (e.g. metric size).

Hydrothermal caverns are known from cave exploreration and studies. The most famous one is Cova des Pas de Vallgornera. This cave seems to result from different karst-forming processes: coastal karstification (mixing water), meteoric water recharge (meteoric karst), together with a probable basal recharge of hypogenic origin (hydrothermal karst sensu largo; Gines et al., 2008; Merino and Fornos, 2010). The latter is supported by hot water resurgence as the thermal spring at Font Sant, located on a fault plane. It discharges water that is sulphate-rich and

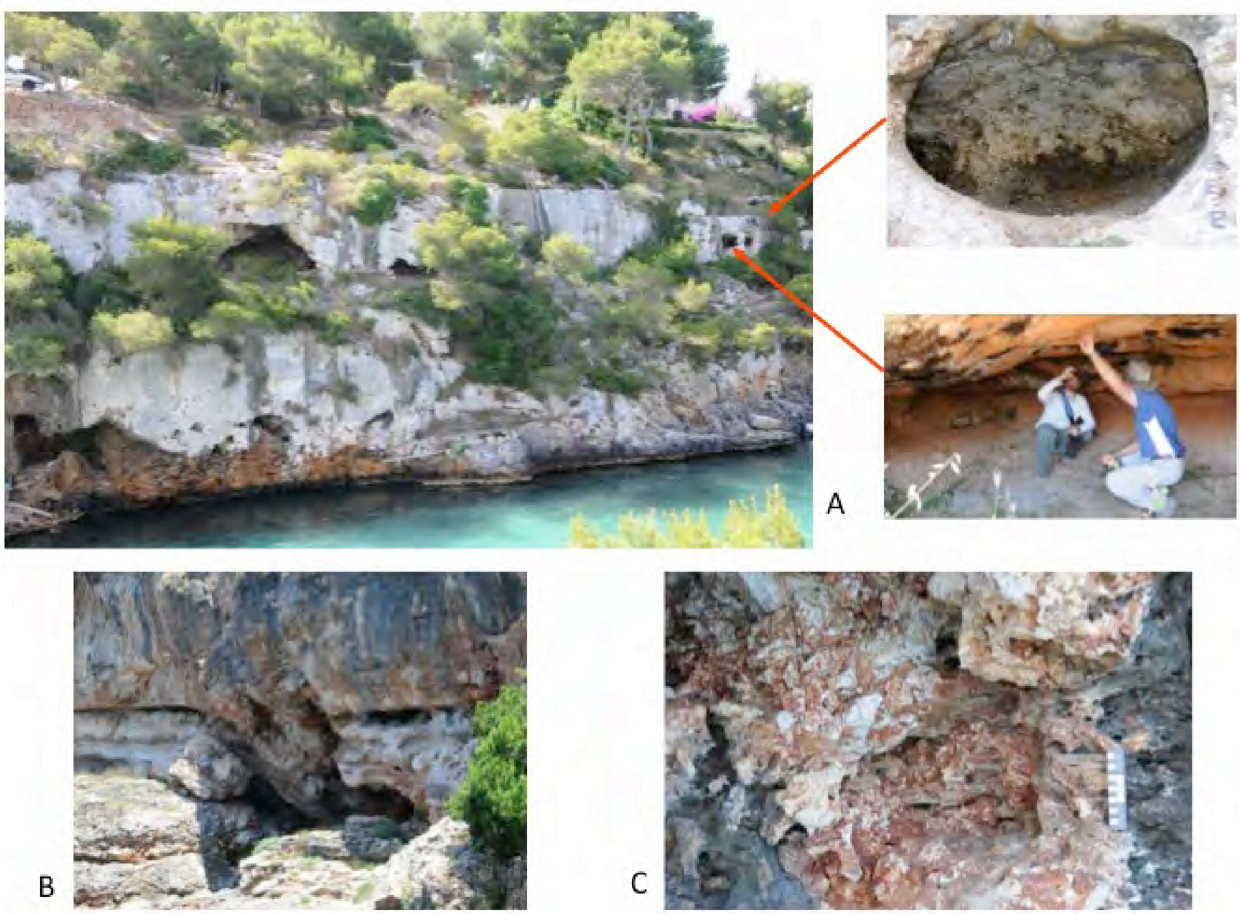

Figure 2. Main karstic features, A) Flank margin caves at Cala Pi, B) Oblique view of a major collapse, 10's m high , Cala Figuera; C) Sponge karst, Cala Bertran.

calcite-rich, with a CO_2 content anomaly and temperature averaging 45-50 °C. Temperatures are in the range of 80-90 °C at 2000m depth (IGME, 2003; Lopez Garcia et al., 2004; Mateos Ruiz et al., 2004).

FAULT AND FRACTURE NETWORK

Mallorca can be considered on the large scale as a horst and graben system that developed across a compressive wedge. NE-SW (N130°) and NW-SE (N50°) structural trends can be interpreted from outcrops as well as satellite imagery in basement and Miocene carbonate platform rocks, and are compatible with the structural trends proposed by Fornos et al. (2012). Thermal anomalies (with hydrothermalism) are rather associated to the NE-SW trending fault trace (N50° trend). Major faults appear to be regularly spaced by 5 to 10 km. The Campos normal fault, trending NE-SW and bounding the Campos basin to the SE, is associated with a thermal anomaly characterized by hot springs known from Roman times at “Font Sant,” about 4 km off the Mediterranean coast. It has been penetrated by a hydrogeological observation hole (S23), intersecting the top of the fault at 14 m (Hebert, 2011). These directions control the development of the Cova Des Pas de Vallgornera. N50° trend is frequently observed as preferential direction for meteoric cave overprint or enlargement of flank margin caves.

POROUS NETWORK QUANTIFICATION – USE OF THE X-RAY COMPUTED TOMOGRAPHY

The CT Scan measures the X-ray density of materials crossed by X-rays. The CT scan image is mathematically reconstructed using the intensities of the transmitted X-ray beam collected at regular increments of rotation around the sample (Ashi, 1997). The intensity of the transmitted X-ray beam is usually expressed as the CT number (Hounsfield scale), i.e., the ratio of the linear attenuation coefficient μ of the material to that of pure water.

Samples

A set of hand specimen collected at Cala Figuera outcrop was analyzed to characterize the porosity in the sponge karst. In order to maximize the results, the hand specimens were cut as cylindrical shapes to provide a better image than an irregular one during acquisition through a medical scanner.

Four wells were selected from the wells cored during the drilling operations conducted by the Bureau of mineral resources of the Mallorca province, namely; MC-2, MC-10, S-19, and S-20. Their respective locations are reported. Along with the CT Scan analyses, a set of conventional core plugs was prepared and measured for calibration.

Virtual Plugging and Virtual Porosity

Virtual plugging consists in sampling a virtual plug of 60 millimetres in length and 39 millimetres in diameter inside the 3D core image. Virtual plugs were taken every 2 centimetres corresponding to 4813 plugs for MC10. With the estimated porosity and X-ray density, a regression curve equation permits to calibrate X-ray density. Using the regression curve equation, virtual porosity can be calibrated, so the curve of virtual porosity of the core can be built every 2 cm using Wellcad©.

Classification of the Porous Network

The next step was to extract and classify the pores by volumetric classes for differentiating, matrix pores, vugs, caverns, etc.; looking for the karst features along the cores. Porosity can be determined by partitioning the voxels (pixel3) belonging to the porosity and the rock matrix using standard image segmentation (Serra, 1982). Density classes are defined by assigning a bin to each grey-level on the image, i.e., determining the grey level range corresponding to pixels having similar properties. The starting point was the x-ray densities computed and the curve of virtual porosity developed for each wells. A specific calibration for the vacuole sizes was devised using test samples with known sizes and associated volumes measured and computed. Then the virtual porosity computed for each core was continuously analyzed based on the vug sizes calibration.

APPLICATION OF THE RESULTS FOR KARST MODELING

Results

Tafonis represent a relatively negligible porosity as they are on the steep flank of the coastline and prone to filling during the next sedimentary phase. Fractures porosity estimate is related to their frequency and as a general rule porosity is low with a maximum less than 0.05%.

Flank margin caves participate as mega pores to the general karstic porosity. The number of sea level changes identified guides their distribution, the measured shapes and the brackish water wedge onto the salt (marine) water intrusion. As a rule of thumb, it decreases away from the coastline. Collapse structures are generated by cave collapse. The related porosity is low compared to the cave porosity. It is similar to breccia porosity. The equivalent porosity can be computed in a first approach using the results published by Loucks (2007). The number and spacing of the collapses reflect the caves distribution, between 20 to 600 m (Robledo, 2002) on 2 sectors studied on the west of Cabo Blanco and on the Eastern coast. The sponge karst dissolution consists of a network of interconnected pores, generally 1–2 cm in diameter with typically sharp smooth to ragged walls, with residual pits and irregular micro-fissures. The amount of dissolution porosity ranges between 6 and 29%, as estimated by CT scan analyses, the larger percentages being more common where the spongy horizons reach maximum thickness.

Each drill core CT-scanned provides a continuous porosity evaluation with the proper calibration performed on conventional core plugs. The extraction and classification of matrix pores, vugs, caverns, provide a porosity estimate for each class that can be directly used for modeling.

Karst Modeling

Karst Modeling for Flank Margin Caves

As expressed above, flank margin dissolution is linked to the paleo-position of sea level at the border of the platform, leading to the definition of a mixing water zone, between seawater and freshwater. This mixing zone is the field of strong dissolution phenomenon leading to flank margin cave to matrix dissolution creation. Dissolution intensity is the strongest near coastline and decreases inland (Labourdette et al., 2007; Labourdette et al, 2013). Areas of dissolution are simulated using truncated gaussian simulations with a proportion defined by coastline position and exponential law decreasing inland. This flank margin dissolution process is a continuum in the size of dissolution features. Vug size and density are simulated with sequential gaussian simulations with exponential variograms with a correlation coefficient for co-simulation with the distance from coastline. This co-simulation allows developing large vugs or caves near the coastline and smaller vugs inland, with a decreasing density. The equivalent petrophysical properties are analyzed to extract deterministic laws linking geometrical properties, porosity and permeability. As for porosity, direct links between permeability and vug density have been found.

Karst Conduits Modeling

This modeling approach leads to horizontally-driven and vertically-driven cave development controlled by all the existing heterogeneities. The resulting complex reservoir is unique and its modeling is achieved through specific geomodeling tools that were internally developed within Total. The software copes with the karst distribution and takes in to account the geological uncertainty associated to that model (Labourdette et al., 2007; Labourdette and Lapointe, 2012; Labourdette et al., 2013; Lapointe and Massonnat, 2008). Conduits are drawn stochastically, crossing the cells of the models. The structural or fracture network is taken into account to generate vertical conduits. The matrix permeability is used to generate horizontal conduit branches. For hydrothermal karst, stochastic seeds are simulated in a defined infiltration zone (base of major stratigraphic layers), at the intersection with the defined fracture network. Vertical conduits are aligned on the fractures. Horizontal conduits are generated when matrix permeability is higher than a defined threshold.

Karst Conduit Petrophysical Properties

Karst conduits are represented in the model as a set of tubes crossing a given cell from one face to the other. During the simulation, conduit radius and conduit number per cells are stored. From these values and cell faces surfaces, equivalent petrophysical properties are calculated using Poiseuille & Darcy's derived laws. The impact of conduits on permeabilities is very strong. Even very small conduits can generate high permeability values; on the contrary, the impact on porosity is weak.

Diffusive Karst Simulation

In the previously defined possible karstified region, the software simulated conduits and associated petrophysical properties. Around these stochastic conduits, an area affected by diffuse karst (matrix dissolution) is defined. Petrophysical properties are simulated in this region, using Gaussian simulations. The permeability is simulated collocated to porosity values with a correlation coefficient.

CONCLUSIONS

Carbonate karst-controlled reservoirs are economically interesting. Mapping of the actual reservoir remains critical for predicting the best porosity-permeability zone distribution and the prognosis of future production well locations, particularly for horizontal or highly deviated wells. The karst modeling workflow applied on actual oil field is supported by the parameters deduced from the outcrop studies of the Llucmajor platform that provide a mandatory set of data. The data set comprises the distribution and sizes of the large caves (flank margin caves, hydrothermal and meteoric caves), the distribution of the sponge karst type with respect to flank margin caves, and its related porosity. The evaluation of the porosity for the diffusive karst as derived from the CT-Scan analysis of the cores.

The 3-D approach leads to horizontally-driven and vertically-driven cave development controlled by all the existing heterogeneities and particularly those related to the tectonic events. The resulting complex reservoir is specific. Its identification and understanding are of paramount importance for reservoir development, particularly in Paleozoic rocks.

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HYPOGENE SPELEOGENESIS AND CO_2: SUGGESTIONS FROM KARST OF ITALY

Marco Menichetti[1]

The carbon dioxide produced in the soil and dissolved in the percolation water is considered as the main agent for karstification in the carbonate rocks. Superficial morphologies and underground caves are product of the corrosion of the limestone, while carbonate speleothems is the other end member of the process.

Hypogene speleogenesis driven by deep seated fluids is the cave formation processes for the main karst systems in the Apennines of Italy. Hydrogen sulfide and endogenic carbon dioxide are the main agents for underground karst corrosion and the soil carbon dioxide plays a secondary rule. The limestone corrosion driven by hydrogen sulfide produces gypsum deposits in caves that could be assumed as the indicator of the hypogene speleogenesis. The action of endogenic carbon dioxide in the cave formation, especially if it operates at lower temperature, is not easy to detect and the resulting cave morphology is not helpful to recognize the cave formation process.

The main sources of carbon dioxide in the underground karst system in the Apennines of Italy can be related to different processes driven by the endogenic fluids emissions. The crustal regional degassing seems to be the prevalent source for carbon dioxide in the karst massifs with the main release in the groundwater. Hydrogen sulfide and methane oxidation, possibly mediated by bacteria activity, are other sources in the buried Cenozoic sediments. Releasing of carbon dioxide along the faults and in the fractures occurring in the carbonate rocks is an important source, especially in the seismically active area. Finally, thermogenic reactions with carbonate rocks are well known as one of the main production mechanism of carbon dioxide released in the atmosphere.

Data from carbon dioxide monitoring in several caves show a relevant contribution of the endogenic carbon dioxide (about 75 %) in the karst system which drives the speleogenesis reactions and shapes the underground morphologies.

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ACTIVE HYPOGENE SPELEOGENESIS IN A REGIONAL KARST AQUIFER: AYYALON CAVE, ISRAEL

Israel Naaman[1,3], Chanan Dimentman[2], and Amos Frumkin[3]

Ayyalon Cave is the longest (2.7 km) and most important known cave in the Ayyalon Saline Anomaly (ASA), Israel (Frumkin and Gvirtzman 2006; Naaman 2011). Located in the inner costal plane of Israel, 21 km from the Mediterranean Sea, it has been truncated by an open quarry in the center of ASA.

The cave comprises two levels of horizontal passages, each with distinct characteristics. The upper level is a network maze cave. The passages have well rounded smooth walls and the ceiling is rich with cupolas and rising cupolas chains. The cave morphology and water properties show that the cave was formed by hypogenic flow of aggressive water that penetrated upwards through vertical shafts and fractures in the bottom of the passages. The findings suggest that the cave was formed in the phreatic zone, beneath the water table, by slow water flow from a deep source into the cave formation zone.

Wall etching is common in the cave, particularly observed around silicified concretions and fossils, which protrude from the etched wall. The etching is often associated with soft, fluffy weathered walls. The soft weathered rock walls and secondary gypsum and calcite deposits in the upper levels indicate condensation corrosion and associated deposition. These processes are attributed to air convection that took place under vadose conditions, after the water table dropped. The nature and distribution of these deposits allowed us to estimate the tracks of the air convection currents. These currents of air convection supplied oxygen to and received heat from the surface of the groundwater and were associated with convection in the groundwater and distinctive dissolution pattern of vertical water flow.

No clear hydrologic connection was observed between the voids and the surface. The observed shafts are completely smooth and rounded. They terminate upward in round cupolas or chert beds, where horizontal galleries are commonly developed. The only clastic sediments within the cave passages are fine clays (apart from breakdown debris).

The largest chamber in the cave is roughly circular in plan, ~40 m in diameter. At the bottom of the cave we have reached the water table of a warm, H_2S-rich aquifer, associated with modern dissolution features. The most notable dissolution features are underwater karren resembling subaerial rillenkarren. These are attributed to the downward part of convective aggressive water flow, descending along the walls

Deformation and collapse features are common in the larger voids of the cave. These include post-cavity movements along bedding planes and fractures, as well as chambers whose ceiling has partly collapsed. Some cave voids have been blocked by collapse and deformation.

In this location, the groundwater of the Yarkon-Taninim Aquifer is warmer and more brackish than the surrounding groundwater, has reduced conditions, and contains hydrogen sulfide.

The surrounding Yarkon-Taninim Aquifer, the western part of Israel's Judea Group aquifer supplies annual average of 360 million cubic meters. Today, this aquifer is endangered due to intensive pumping. The fall of the water table in Ayyalon Cave impacts the cave and particularly its ecosystem (below). The replenishment zone of the aquifer extends along the western slopes of Judea and Samaria mountains, while most pumping wells spread along the western foothills of the Shefela region, where the aquifer is commonly confined. The aquifer rocks are predominantly Cretaceous carbonates interbedded with thin marl layers, with a total thickness of about 800 m. Most wells pump from its shallow part, termed the "upper sub-aquifer," consisting of late-Cenomanian and Turonian rocks, while the Albian and lower-Cenomanian rocks, the "lower sub-aquifer," have been hardly drilled, so its properties are less known. Ayyalon Cave is developed in the upper sub-aquifer within Turonian limestone of Bi'na Formation.

The groundwater exposed in the cave is highly representative of the ASA, with chloride concentration of 524 mg/l and total dissolved solids of 1371 mg/l. The sulfide concentration is 5 mg/l and the water temperature (29°C) is 5°C higher than the average temperature in the aquifer. In some measurements we did not identify presence of dissolved oxygen in the water, but its long-term presence is obvious, possibly at the uppermost water layer, as it is crucial for the aquatic organisms. The water table in the area dropped by 13 m since 1951. However, there were no significant changes in the water chemical composition, and no significant correlation between the water properties and the groundwater level. Comparison of ionic ratios of the water in the cave and water wells in the vicinity reveals a mixture of two different salinity sources of water, and shows that the water in the Ayyalon Cave is closest to the saline end-member.

The lower level is characterized by large collapse halls and debris piles covering the bottom of the passages and halls. The

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morphology of this level indicates speleogenesis under phreatic conditions. Until recently, speleogenesis activity occurred on most of this level surface, but anthropogenic over-use of the regional aquifer since mid 20th century lowered the water-table to the point that this level has almost dried out. Consequently, the "wetland" was reduced to a fraction of its original size.

The Ayyalon Cave has unique ecosystem. Lacking sunlight and a constant supply of organic material, the ecosystem is based on chemoautotrophic primary production, by *Beggiatoa-like* bacteria that create a bacterial mat on the water surface. The isotopic composition of carbon and oxygen in the fauna and the bacteria shows that the latter are the energy source for the system. The cave groundwater contains new species of eyeless, pigmentless stygobitic crustaceans. Troglobitic terrestrial arthropods were found in the adjacent area within the cave.

The faunistic assemblage consists of six endemic stygobiont and troglobiont crustaceans and other arthropods. Two species are still under unclear status (Por et al., 2013). Some species, such as the scorpion *Akrav israchanani* apparently became extinct, probably due to the massive drop of water level in the cave and resulting reduction in primary production at the water surface.

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HYPOGENE VS EPIGENE CAVES: THE SULFUR AND OXYGEN ISOTOPE FINGERPRINT

Bogdan P. Onac[1]

The classical epigene speleogenetic model in which CO_2 is considered the main source of acidity has been challenged over the last three decades by observations that revealed cave passages unrelated to groundwater drainage routes and surface topography. Most of these passages show unusual morphologies, such are cupolas, floor feeders (i.e., inlets for deep-seated fluids), and huge irregular-shaped rooms that terminate abruptly, and often a rich and diverse mineral association. A hypogenetic speleogenetic pathway was proposed for this group of caves.

The presence of abundant gypsum deposits in caves with one or more of the passage morphologies listed above, have prompted scientists to suggest a new theory (i.e., sulfuric acid speleogenesis, SAS) of cave development. In the hypogenic SAS model, the source of acidity is the sulfuric acid produced by oxidation of H_2S (originating from sulfate reduction or petroleum reservoirs) near or at the water table, where it dissolves the limestone bedrock and precipitates extensive gypsum deposits. SAS is now thoroughly documented from numerous caves around the world, with the best examples coming from the Guadalupe Mountains (NM), Frasassi caves (Italy), selected caves in France, Cueva de Villa Luz (Mexico), and Cerna Valley (SW Romania).

To date, discrimination between epigene and hypogene speleogenetic pathways is made using cave morphology criteria, exotic mineral assemblages, and the predominantly negative $\delta^{34}S$ values for the cave sulfates. This presentation highlights the role sulfur and oxygen stable isotope analyses have in discriminating between epigene and hypogene caves.

Based on a number of case studies in caves of the Cerna Valley (Romania), we found that relatively S-depleted isotopic composition of cave minerals alone does not provide enough information to clearly distinguish SAS from other complex speleogenetic pathways. In fact, $\delta^{34}S$ values of SAS by-products depend not only on the source of the S, but also on the completeness of S redox reactions. Therefore, similar studies to this are needed to precisely diagnose SAS and to provide information on the S cycle in a given karst system.

Integrating cave mineralogy, passage morphology, and geochemical studies may shed light on the interpretation of polygenetic caves, offering clues to processes, mechanisms, and parameters involved in their genesis (sulfate-dominated).

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PER ASCENSUM CAVE MORPHOLOGIES IN THREE CONTINENTS AND ONE ISLAND, INCLUDING PLACES WHERE THEY SHOULDN'T OCCUR

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Hypogene or per-ascensum, whatever you prefer to call them, caves that form from the bottom up have a great range of patterns in plan, large cavity morphology and an expanding, but specific suite of speleogens that distinguish them from fluvial caves formed by descending surface water. Once thought to be rare and unusual, caves or sections of caves with plans, large cavities and suites of "hypogene" speleogens are turning up in situations traditionally thought to have fluvial or even glacial origin. The role of condensation corrosion in the formation of cavities and speleogens remains controversial, but surprisingly some insights may come for processes in salt mines. Phantom rock formation and removal and similar processes involving removal of dolomitized bedrock, de-dolomitized bedrock, and almost trace-free removal of palaeokarst raise problems of both temporal relationships and of how to distinguish between the outcomes of recent and ancient processes. The presence of "hypogene" speleogens in both gneiss and marble caves in Sri Lankan of unclear origin adds to the complexity. Back in the early 1990s, before hypogene caves were de-rigour, workers such as David Lowe were puzzling about speleo-inception, how caves begin. Perhaps the rare occurrences of solution pockets in joints in obvious fluvial caves, such as Postojna Jama, are indicating that many more caves than we imagine are actually multi-process and multiphase and that "hypogene" processes of various types are significant agents of speleo-inception.

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HELIUM ISOTOPES AS INDICATOR OF CURRENT HYPOGENIC KARST DEVELOPMENT IN TAURIDS KARST REGION, TURKEY

N. Nur Ozyurt[1] and C. Serdar Bayari[1]

Hypogenic karst development by means of the aggressiveness of hydrothermal fluids driven and fed by mantle heat and mass flux is a known phenomenon. However, in cases when hydrothermal fluid cools down upon thermal conduction in the near-surface environment and is diluted by near-surface cool groundwater, evidences of this phenomenon may be erased completely. Recent data on the isotopes of helium dissolved in cool karst groundwater samples collected from three different karst aquifers in Turkey suggest an apparent mass flux from mantle, as well as from the crust. In the cases considered, helium content from the mantle increases with the increasing age of groundwater. All cases are located nearby the suture zones which may be easing the upward heat and mass flux. Despite sampling difficulties and high analysis costs, helium isotopes dissolved in cool karst groundwater seem to be useful tool to detect the current hypogenesis at the depths of karst aquifers.

INTRODUCTION

Hypogenic karst development that occurred in the past can be detected by using its characteristic morphological evidences kept in caves or on the surface. However, current hypogenic karst development may not be easily detected if the process does not have clear indications like an apparent hydrothermal activity on the surface.

According to an early definition, hypogenic speleogenesis is attributed to uprising hydrothermal fluids in which "*aggressiveness has been produced at depth beneath the surface, independent of surface or soil CO_2 or other near surface acid sources*" (Palmer, 1991, in Klimchouk, 1999). Apart from the source of aggressiveness, this type of hypogenic karst development also requires a heat source. The heat transfer to deep-seated fluid reduces its density and increasing buoyancy forces the fluid to rise toward the surface. Source of the heat may be radiogenic, the mantle itself, or any magmatic or volcanic intrusion in the crust. On the other hand, the indicators of current hypogenic speleogenesis may be weakened or wiped out during the upwelling. For example, the initial heat may be reduced to the level of near-surface environment, and characteristic element (e.g. Li, B, Br) contents may be reduced to background levels upon mixing with cool-shallow groundwater. In such cases, isotopes of helium dissolved in karst groundwater may help to determine whether the flow system is under the influence of current hypogenic karst development or not. In the following, we briefly explain the signatures of different helium isotopes sources and then, present helium isotope data collected from Konya Closed Basin (KCB), Kırkgöz Karst Aquifer (KKA) and Aladag Karstic Aquifer (AKA), all belonging to the Taurus karst belt of Turkey (Fig. 1).

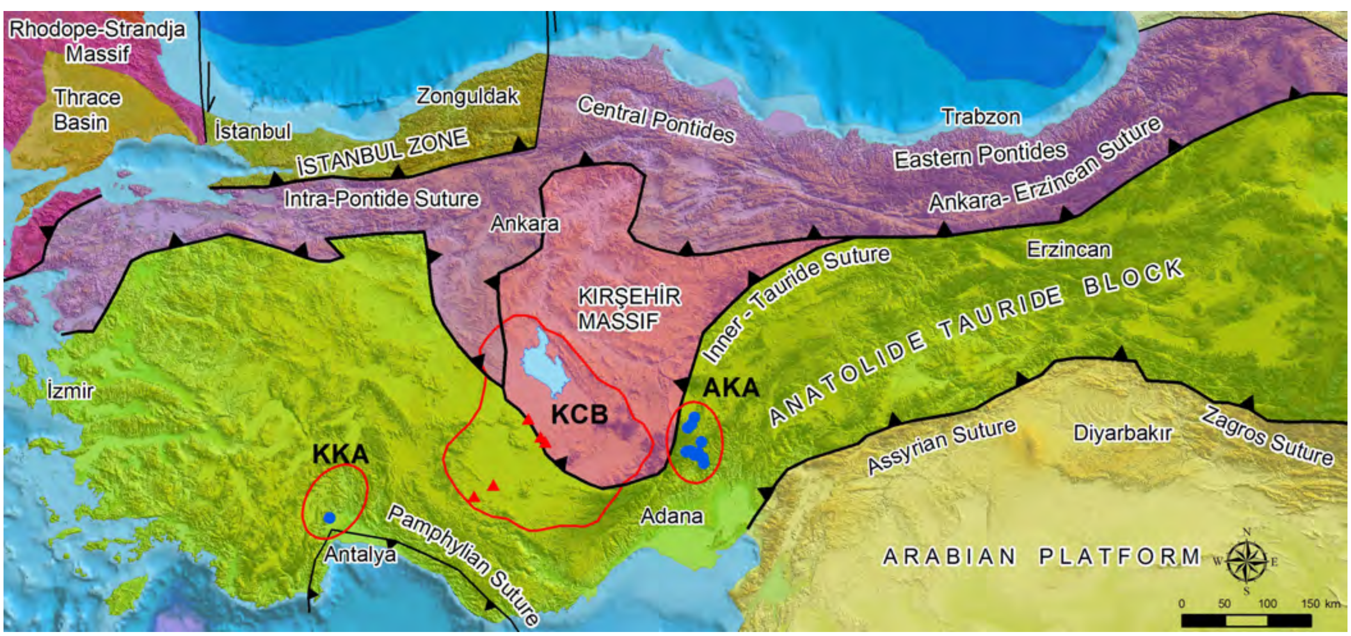

Figure 1. Locations of helium isotope sampling sites and the suture zones in Turkey (modified after Okay and Tuysuz,1999).

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HELIUM ISOTOPES AS INDICATOR OF HYPOGENIC KARSTIFICATION

Noble gases (He, Ne, Ar, Kr, Xe) are rare in the earth. Since they are not involved in chemical reactions, their isotopic signals can be used to identify various natural processes. Using noble gases in geochemistry started in early 1900s, mostly for geochronological applications (Ozima and Podosek, 1983). Later on, in 1970s, following the establishment of the data on their solubility in water, they have been used in hydrogeological and hydrogeochemical studies. Today, noble gases are widely used for recharge temperature estimation, age-dating (based on tritium - tritiogenic helium-3 ratios) and aquifer characterization. Among all noble gases, helium isotopes are particularly useful in geological and hydrogeological applications because of their characteristic signals of the sources of atmosphere, crust and mantle. The helium element practically has two isotopes, Helium-3 (He-3) and Helium-4 (He-4). Other isotopes of helium have lifetimes less than a second. Helium isotope ratios (i.e. He-3/He-4) have been presented generally by using "R" (ratio) notation in the literature. Atmospheric $^3He/^4He$ ratio (Ra) is 1.38×10^{-6}. Because of the continuous mixing of the atmosphere, this number is spatially fixed. Measured helium gas ratios are used as normalized figures against the atmospheric ratio, which is shown as R/Ra. Typical R/Ra values for atmosphere, crust and mantle are 1, 0.05, and 8, respectively (e.g. Ozima and Podosek, 1983). Sources of He-3 in groundwater are atmosphere, fission of Li-6, which is a product of U-Th decay in the crust, escape from mantle, and decay of tritium isotope. Sources of He-4 in groundwater are atmosphere and U-Th series decay in the crust. Mantle source of He-4 is negligible.

Applications of noble gases in karst aquifers are pretty rare because of possible signal modification by the excess air. Turbulent flow conditions in the karst aquifer causes the entrapment of air bubbles whose dissolution results in elevated gas concentrations compared to the groundwater, which is in equilibrium with atmosphere at the time of recharge. However, whether a groundwater sample involves excess air can be easily detected by using the neon to total helium concentration ratio, which is fixed in a water sample that is in equilibrium with the atmosphere at a given temperature.

HELIUM ISOTOPE OBSERVATIONS FROM TAURIDS KARST REGION

A summary of helium isotope data and hypogenic karst features of the water samples collected from Konya Closed Basin (KCB), Kırkgöz Karst Aquifer (KKA) and Aladag Karstic Aquifer (AKA) are presented in Table 1. Detailed assessment of the data is presented in the following sections.

Konya Closed Basin

The Konya Closed Basin (KCB), located in central Anatolia (Turkey), is one of the major (53,000-km^2) endorheic basins in the world. KCB is semi-arid; hence, the groundwater is the only dependable water resource. KCB comprises of southern Konya and northern Salt Lake (Tuz Golu, in Turkish) sub-basins (Bayari et al., 2009a, b), which are separated by a roughly E-W trending plateau. The plateau is abounded with obruks (Turkish name of gigantic collapse dolines), which are hypogenic karst landforms. According to Bayari et al. (2009a), "Presence of volcanogenic elements (i.e. Li and F) and remarkably high dissolved carbon dioxide (log $PCO_2 = 10^{-1}$ atm) in fresh groundwater, hydrothermal springs with elevated He contents (R/Ra = 4.77), highly enriched carbon-13 isotopic composition of total dissolved inorganic carbon (^{13}C_TDIC = −1.12‰ V-PDB) in the regional groundwater and presence of widespread carbon dioxide discharges, constitute apparent evidence for the hypogenic fluid migration into the Neogene aquifer where enhanced dissolution due to mixing between the shallow-fresh and deep-saline groundwater gives rise to obruk formation."

A simplified tectono-stratigraphic sequence starts with Paleozoic-Mesozoic, rich in carbonate rocks forming the Taurus Mountains at the south and the most of the central plateau. On top of this, carbonate, detrital and evaporite rocks of Paleogene series are located. Overlying Neogene is made up mainly of lacustrine carbonate rocks. The sequence is completed by Plio-Quaternary lake deposits that are practically impermeable and cover the central parts of the sub-basins.

The Neogene lacustrine carbonates exhibit well-developed karst that constitutes the main freshwater aquifer of the basin. The aquifer is confined by the overlying Plio-quaternary impermeable cover and is very productive. In between the sub-basin, through the central plateau, the aquifer becomes unconfined because of the absence of Plio-quaternary cover.

Taurus Mountain Range is the main recharge area of the KCB. Groundwater flows through the Neogene aquifer toward the terminal Salt Lake. Radiocarbon ages (ranging from recent to over 40,000 years BP) of groundwater increase systematically from recharge area to Salt Lake (Bayari et al, 2009b). Typical karstic features of hypogenic karst development in KCB are huge collapse dolines and travertine cones (Bayari et al, 2009a-c). Collapse dolines are upside down truncated cone shaped and depths and top radius vary between 34 to 225 m and 100 to 300 m, respectively (Bayari et al., 2009b). Many of those collapse dolines constitute groundwater windows. Formation of the collapse dolines are directly linked to geogenic (i.e. mantle and crust) CO_2 (Bayari et al., 2009-b-c).

Groundwater in the Neogene aquifer were sampled from 10 deep wells located along two transects extending from the recharge area to Salt Lake in August 2011 to determine noble gas composition. Samples were taken in copper tubes to prevent contamination by air and have been analyzed at Isotope Hydrology Laboratory of the IAEA. Measured He-3 and He-4 concentrations vary between 10^{-13} to $10^{-12}cm^3$ STP/g and 10^{-7} and 10^{-6} cm^3 STP/g, respectively. Except one sample (KCB7),which is very

Table 1. R/Ra ratios and hypogenic features of selected karst aquifers in Turkey.

	R/Ra range Number of sampling points	**Hypogenic Karst Features**
Konya Closed Basin	0.88-2.13 10 water samples (well)	Obruks (giant collapse dolines), travertine cones
Kırkgöz Karst Aquifer	1.01-2.60 5 water samples (5 springs)	Giant cave system, travertine deposits
Aladag Karstic System	0.82-1.14 12 water sample (6 springs)	Travertine deposits, cave morphology

close to the recharge area, all samples have remarkable amount of Helium-3, which has been released from mantle (Fig. 3). Quantifying magnitude of mantle He-3 is difficult and for most cases is impossible because of the mixture of gases from various sources. He-3 content of groundwater along both transects reach maximum values just before the Obruk Plateau where the aquifer becomes unconfined. Beyond this point, gas composition of groundwater diluted by recharge from plateau and/or degassing of unconfined groundwater (Fig. 4). Although, any R/Ra ratio greater than 1 is associated with mantle gas contribution, simultaneous increase in He-4 content also indicates presence of crustal He-4 production along the groundwater flow path of KCB. R/Ra values up to 4.77 have also been observed in thermal groundwater samples at the NE boundary of the KCB (Gülec et al., 2002).

Almost all collapse dolines are located along the suture zone between Anatolide-Tauride and Kirsehir Massif continental fragments (Fig. 1 and Fig. 2). It seems that this suture zone provides a preferential route for heat and mass escape from mantle.

Kirkgöz Karst Aquifer

KKA is located at the SW part of the Taurus Mountains. KKA is drained mainly by the Kirkgöz group of springs whose mean annual discharge is 18.5 m^3/sec. KKA's spring zone is located along the foot of mountain where carbonate rocks with well-developed karst overlain tectonically by impermeable ophiolitic rocks at elevation of about 300 m asl. Kirkgöz springs form a pond of 7 hectares. The world's largest (~ 650 km^2) freshwater travertine deposit has been formed by these spring waters since the Pliocene. Total volume of the travertine is 90,109 m^3 for 25% porosity and about 200 m mean thickness. KKA hosts many caves but one of them is remarkable because it hosts world's largest underwater cavity, called The Stadium (Kincaid 1999). The cavity measures 100 m by 60 m by 45 m and has the volume of 270,000 m3 (Kincaid, 1999; Kincaid and Jablonski, 1996). None of the springs (with similar hydrogeologic setting and discharge rate) throughout the Taurus Mountain Range have formed such large travertine deposit and host submerged cavities that are comparable to the Stadium. Such a large dissolution features in the discharge zone of KKA and the associated travertine deposit requires a vast amount of aggressiveness which appears to be associated with hypogenic sources.

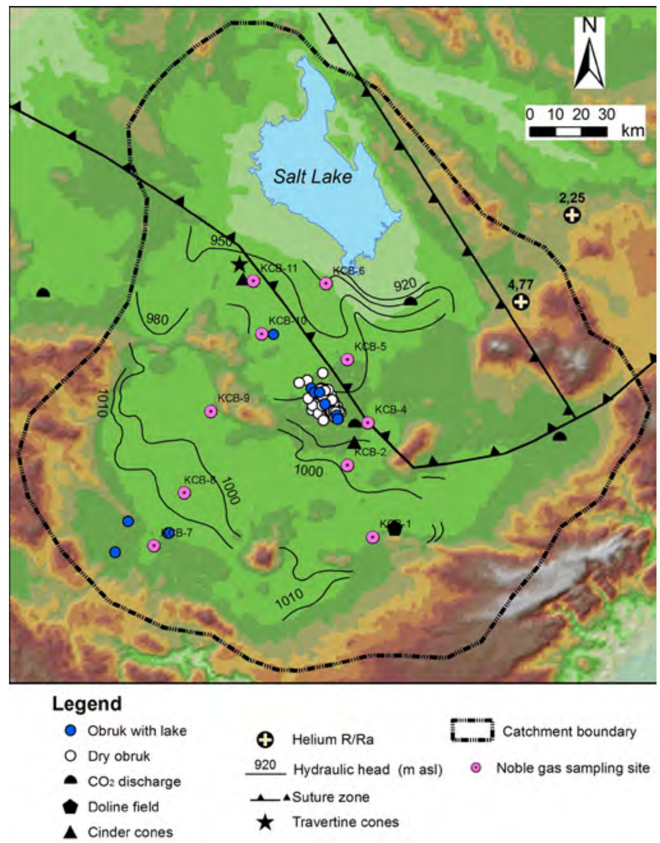

Figure 2. Location map of sampling sites and collapse dolines of KCB (modified after Bayari et al., 2009b).

The Kirkgöz group of springs were sampled at 5 different discharge spots to determine the noble gas content of groundwater in 1997, and the samples analyzed at Noble Gas Laboratory of University of Utah (Nativ et al., 1999). Measured R/Ra values of those samples vary between 1.01 and 2.60. The R/Ra figure of 2.60 is the highest ratio measured so far in the karst springs in Turkey. This high R/Ra ratio apparently indicates heat and mass transport from mantle to deep-seated fluids underneath the KKA. Like KCB, KKA is located in close proximity of the Pamphylian suture zone, which may be easing heat and mass escape from mantle. The epicenter depths reaching 125 km around the KKA is an apparent indicator of deep crustal activity, which is very likely to be associated with heat and mass escape from the mantle.

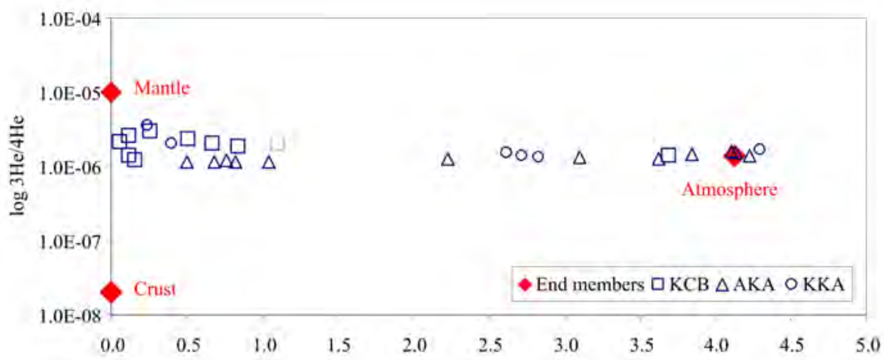

Figure 3. Origin of dissolved helium in karst groundwater samples.

Aladag Karst Aquifer

AKA covers an area of 1,065 km^2 in the Aladaglar Mountain Range, located in south-central Turkey. The mountain range, extending between 400 m asl and 3750 m asl, hosts one of the deepest karst aquifers in the world (Ozyurt, 2005). The karst of the entire range is very well developed so that, despite extensive glacial scouring of the karst surface during Quaternary, more than 250 caves have been found during 7-years cave survey (Klimchouk et al., 2006). The range includes numerous carbonate-hosted lead-zinc deposits that are initially sulfidic. Almost all caves and debris-filled karst cavities encountered in mine galleries are abounded with the evidences of hydrothermal karst development.

The AKA composes two carbonate nappe units that are separated by ophiolitic mélange. Limestone and dolomite belonging to Devonian-Cretaceous and Triassic-Cretaceous periods are the dominant rock units in the carbonate nappes. Main drainage of AKA is towards to Zamanti River that crosses mountain range from north (at 1000 m asl) to south (at 400 m asl) (Ozyurt and Bayari, 2008). Along the groundwater flow paths to Zamanti River, there are several groundwater drainage routes arriving at specific springs. Some of the springs deposit travertine in the form of natural bridges over the river (Bayari, 2002).

Six major springs in AKA were sampled to determine the noble gas contents of karst groundwater from 2002 to 2004 at various seasons. A total of 12 samples have been analyzed at the Noble Gas Laboratory of University of Utah (Ozyurt, 2005). R/Ra values of the samples were found to range from 0.82 to 1.14. These ratios are much lower than those observed in KCB and KKA. However, values above 1 indicate presence of mantle helium, which seems to be diluted very much by the atmospheric contribution. Moreover, short mean residence time of karst groundwater in the AKA (maximum of 20 years) compared to that in KCB (up to 40,000 years BP) and KKA (around 120 years) indicates that its groundwater has more limited circulation depth and shorter time to collect more deep-seated gas. Among all, 3 samples of AKA exhibits strong gas contribution from atmosphere and even one sample is likely to have excess air (Fig. 3). Like other aquifers, AKA is also located at close proximity of Inner-Taurids suture between the Kirsehir Massif and Anatolide-Tauride continental fragments. Part of the suture at the western part of AKA is an active strike-slip fault along which there exist several thermal springs and volcanic vents which were active at around 253 AD.

CONCLUSIONS AND OUTLOOK

Noble gas data obtained from the groundwater of three different karstic systems in the Taurids karst range of Turkey suggest an apparent noble gas contribution from the mantle and crust. Such

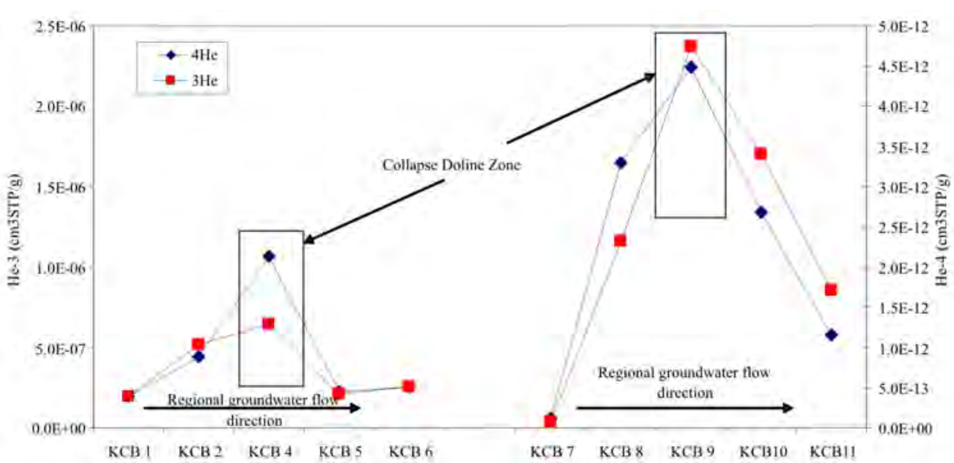

Figure 4. Spatial variation of dissolved helium isotopes along the regional flow path in KCB.

a gas upwelling from mantle to the surface of the crust should be a part of the upward fluid migration, which is driven by the elevated buoyancy of deep-seated fluid by the mantle heat flux. Presence of mantle helium in cool karst groundwater suggests a current mass contribution from the mantle even though other signals of deep-seated fluid contribution are not clearly visible. Despite sampling for noble gas composition is difficult and the analyses are expensive, similar studies should be conducted in the future on cool karst groundwater system located elsewhere to verify the results of this study.

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EVOLVING INTERPRETATIONS OF HYPOGENE SPELEOGENESIS IN THE BLACK HILLS, SOUTH DAKOTA

Arthur N. Palmer[1] and Margaret V. Palmer[1]

The origin of caves in the Black Hills has long been debated. Their history is long and complex, involving early diagenesis, meteoric karst (now paleokarst), deep burial, tectonic uplift, and, finally, enlargement of previous voids to the caves of today. The final stage is usually the only one recognized and is the topic of this paper. Genetic hypotheses include artesian flow, rising flow (preferably thermal), diffuse infiltration, and mixing of various water sources. The last process best fits the regional setting and water chemistry.

INTRODUCTION

The Black Hills were formed by Laramide uplift between 65 and 30 Ma. The central peaks consist of Precambrian igneous and metamorphic rock, and the truncated edges of overlying sedimentary strata dip off to the sides, forming concentric ridges and valleys around the central mountains. Caves are located in the Madison Formation (limestone and dolomite) of early Carboniferous age (Mississippian in North America). The largest caves (Wind and Jewel) are dense fissure networks with several levels concordant to the strata, which locally dip about 5 degrees away from the central uplift.

Their main (final) phase of cave origin has long been debated. Most of the accepted hypotheses involve hypogene processes. They have been attributed by various researchers to (a) artesian flow, (b) rising water, probably thermal, (c) enlargement of relict paleokarst, (d) diffuse recharge through the basal sandstone of the overlying Minnelusa Formation, and (e) mixing of two or more water sources. Each model is examined below vs. local cave morphology, geology, hydrology, and geochemistry. There is evidence for each, some of it spotty, but only the last process (e) survives the scrutiny. Our own interpretations have evolved considerably over the past 35 years.

CAVE MORPHOLOGY

The caves occupy only the upper 100-150 m of the limestone. They are located only beneath the thin up-dip edge of the Minnelusa Formation, which consists mainly of quartz sandstone (Palmer and Palmer, 1989; Wiles, 2005). Updip from the edge of the Minnelusa the caves become very sparse and disappear a short distance away. They are also very sparse beneath incised valleys that have cut through the Minnelusa into the limestone. Passages beneath the limestone do not terminate in breakdown,

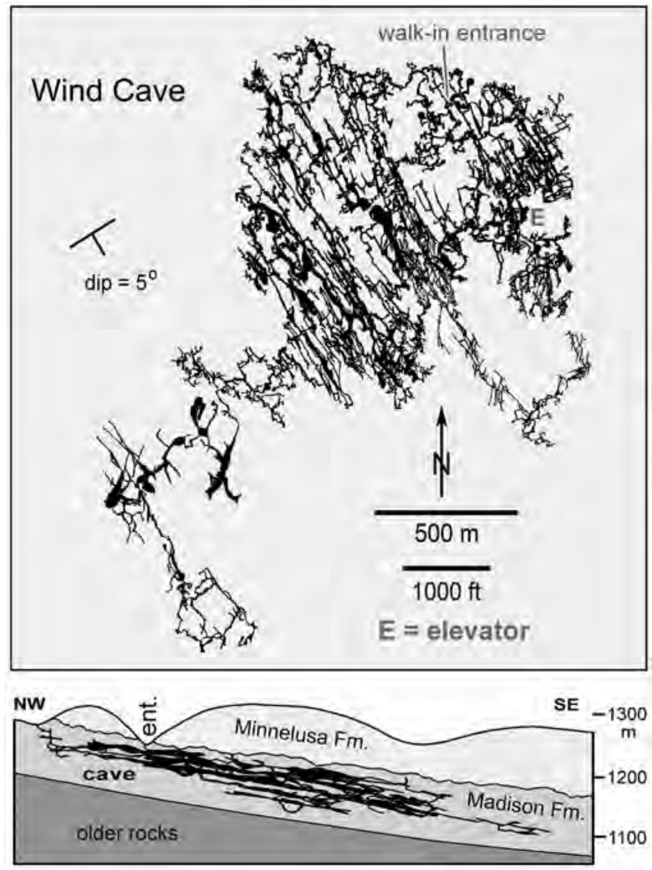

Figure 1. Plan view and simplified geologic profile of Wind Cave, Wind Cave National Park, in the southeastern Black Hills.

but instead simply stop or become very narrow. The caves contain no significant vadose passages. Instead they give the impression of being entirely phreatic, with many domed ceilings and cupolas. Most of these, however, appear to have formed by condensation corrosion via convection cells, a process that continues today.

The caves do not extend beneath the areas of thick overlying Minnelusa or indefinitely below the water table. The water table is reached in only a few places in Wind Cave, with no hint that the caves continue to depth. Jewel Cave does not reach the water table at all. The caves reach the top of the limestone in only a few places where irregularities in a paleokarst surface extends downward. (See Sando, 1988, for a regional description of this paleokarst.) The only direction in which they seem unlimited is along the strike of the beds. The caves are essentially “floating” in the Madison, with almost no contact with boundaries.

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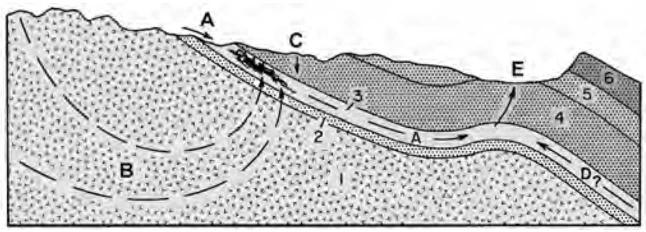

Figure 2. Profile of a typical cave in the Black Hills, showing potential paths for the water flow that formed it. 1 = Precambrian metamorphic and igneous rocks. 2 = Deadwood Sandstone and other strata (mainly Cambrian). 3 = Madison Limestone (Mississippian, lower Carboniferous). 4 = Minnelusa Formation (upper Carboniferous, Pennsylvanian – lower Permian). 5, 6 = overlying strata, mainly shales and sandstones (Triassic–Cretaceous). A = water entering the outcrop area of the limestone (autogenic seepage, allogenic streams). B = rising water, possibly thermal, recharged from the higher hills. C = diffuse recharge through the lower beds of the Minnelusa Formation (mainly sandstone, but with carbonate and sulfate interbeds). D = up-dip flow from surrounding lowlands (unlikely, because it would appear to be against the hydraulic gradient). E = major springs around the perimeter of the Black Hills, where confined water in the Madison passes upward along faults and breccia in the upper Minnelusa and overlying strata (see Rahn and Gries, 1973).

HYPOTHESES OF CAVE ORIGIN

Artesian Flow

An artesian origin was favored by many early studies, because the Madison is a well-known semi-confined aquifer overlain by low-permeability rocks (Tullis and Gries, 1938; Deal, 1962; Howard, 1964). Water rises in the valleys around the Black Hills by rising upward along faults and breccia zones. Up-dip from the major caves are broad valleys, drained today only by small under-fit streams, which indicate significant sources of recharge in the past (Palmer and Palmer, 2008). Such recharge points could have provided the flow necessary to form the caves. However, an artesian origin would produce caves that extend indefinitely down the dip, with their maximum size in the recharge area and diminishing size in the direction of flow. Network patterns have long been associated with artesian flow, but when examined in detail this relationship has little hydrochemical merit (Palmer, 1991). The caves do not extend far in the up-dip or down-dip directions. They cease where the overlying Minnelusa is thick or absent.

Thermal Hypogene Hypotheses

An origin by rising thermal water has gained great favor by the discovery of thermal minerals (quartz crystals, crystalline hematite) and calcite with highly negative oxygen isotopes ($\delta^{18}O$) down to -21‰ (White and Deike, 1962; Bakalowicz et al., 1987). Cupolas and other features resemble those of known thermal caves, as in Budapest, and boxwork veins resemble those of mining areas (Bakalowicz et al., 1987; Ford, 1989). Thick calcite crusts (up to 15 cm) suggest deposition during the last phases of thermal speleogenesis, when rising water approached the water table, degassed, and became supersaturated with calcite. These are strong arguments. However, the thermal minerals, and those with lowest $\delta^{18}O$, formed in palokarst voids during 200 My of deep burial prior to mountain uplift and were intersected by the main cave enlargement. The thick calcite crusts overlie bedrock with deep subaerial weathering, and also cover many older vadose speleothems. They do not relate to the cave origin per se. The lightest oxygen isotopes are those of a 1-cm-thick crust of white scalenohedral calcite, which lines only the remnants of paleokarst voids and was clearly formed during deep burial prior to mountain uplift (Palmer and Palmer, 2008). No dates are available for them because their U content is too low – typical of deep-seated deposits remote from redox boundaries. However, the thickest calcite crusts show U/Pb dates of 25-14.7 Ma (Victor Polyak, in Palmer et al., 2009). These coincide with the peak thickness of the Oligocene White River sediments, which covered springs in the outskirts of the Black Hills and were removed by well-documented mid-Miocene erosional entrenchment (Harksen and Macdonald, 1969; Gries, 1996). So the main cave origin pre-dates the late Oligocene and has been inactive since, except for minor vadose corrosion and deposits. Younger calcite crust has U/Th dates of <300 ka (Ford et al., 1993; Paces et al., 2013). Its fluctuations correlate with late Pleistocene climate variations, and with aggradation and entrenchment of nearby surface rivers.

Finally, the caves do not extend down-dip, and they do not reach the base of the limestone anywhere. Also, they reach the top of the limestone only in a few scattered areas, which are mainly choked with paleokarst fill. This shows that the caves could not have formed by water rising across the strata. Water rising concordantly up-dip through the Madison is a possibility, but it cannot explain why the caves terminate so abruptly both down-dip and up-dip. Also, this water movement would have been against the regional flow pattern, which appears to have been similar to that of today (see reference to thick calcite crusts above). Finally, the clustering of caves beneath the thin Minnelusa cap cannot be accounted for, when we consider that unimpeded flow through nearby uncapped Madison would have been readily available. However, there will always be some enthusiasm for a thermal hypogenic origin, despite formidable limitations.

Thermal Alteration

A relation to late Eocene thermal mineralization is possible, and a magnetic anomaly in the Precambrian below Wind Cave may be pertinent (Kleinkopf and Redden, 1975; Hildenbrand and Kucks, 1985). Many aspects of mineralization in the northern Black Hills resemble those of the boxwork zones in the caves. However, there is strong evidence that the boxwork and other suspect features in the caves are of early diagenetic origin and involved Mississippian sulfate-carbonate interactions, because they do not extend into the paleokarst zone and are covered by deep-burial thermal minerals that pre-date tectonic uplift. Even if the main phase of cave origin has no relation to the Eocene mineralization, the influence of mineralizing fluids should be considered. There is evidence that anoxic fluids entered the caves during the Cenozoic, as shown by Fe and Mn oxides along minor faults.

Enlargement of Paleokarst Voids

Enlargement of paleokarst voids has clearly taken place, because the caves intersect many solution pockets lined by deep-burial minerals, including the white scalenohedral calcite (Palmer and Palmer, 1989). The caves are also concentrated around zones of paleokarst, which includes relics of both early diagenetic features and meteoric karst from dissolution of the Madison prior to deposition of the overlying Minnelusa. Nearby canyons in the Madison encounter a few paleo-pockets, but no major caves – definitely not the great density of passages seen in the caves. The diagenetic features (Mississippian) appear to have formed along zones of former anhydrite, as they include much carbonate breccia, calcite veins, early solution voids, and anhydrite inclusions. The meteoric paleokarst (Mississippian-Pennsylvanian) concentrated around the diagenetic features; and the final cave origin concentrated on enlarging both sets of the earlier voids. But the real question centers on the nature of the water that enlarged these voids to form the present cave.

Diffuse Infiltration

Diffuse recharge through the basal Minnelusa Formation is mildly attractive because many network mazes in other regions cluster beneath thin sandstone caps. However, the caves reach the sandstone/ limestone contact in only a few places, and the upper levels that do reach the contact are sparse irregular rooms, rather than fissure networks. Still, there is some recent support for this idea: Michael Wiles (Jewel Cave, personal communication, 2013) envisions water passing laterally through the Minnelusa, with local loops down into the Madison and then back up again. It is difficult to envision a distribution of hydraulic head that would achieve this flow pattern.

Mixing Processes

Mixing of two or more water sources of contrasting chemistry applies to seacoast aquifers, but does it also apply in the Black Hills? The model of Bögli (1964) and other authors is pertinent here, where two waters of contrasting CO_2 content mix to produce an undersaturated solution. Consider diffuse recharge through the thin sandstone base of the Minnelusa combined with recharge from the outcrop area. Both are active today, but neither contributes significantly to cave enlargement. Most of this paper deals with this model because it seems valid and yet is so seldom discussed. If these two sources were able to mix before the water table dropped, they could have produced the cave in the following way:

Recharge through the sandstone encounters interbedded carbonate rocks near the Minnelusa/Madison contact, and eventually the Madison itself. In closed conditions, with no access to the cave air or soil air, the water would reach calcite saturation at a very low value (~12-15 mg/L) and with extremely low $P\text{CO}_2$ (<0.0001 atm). Meanwhile, the water entering from the outcrop area would also have been close to saturation because of low flow rate and lengthy contact with the Madison carbonates. It would have approached calcite saturation at about 200 mg/L with $P\text{CO}_2$ of roughly 0.005 atm. A mixture of both saturated waters would produce an undersaturated solution capable of dissolving limestone. Mixing and dissolution would have focused on the old paleo-voids, because they offered the highest permeability. The result is what we see in the caves today, at least where the younger calcite crusts are absent.

This hypothesis may seem absurd, but the two waters involved in this process are still present in greatly diminished form, especially in Wind Cave. Trickles of water still enter through the limestone outcrop, and moisture still seeps through the sandstone into the limestone and eventually through the limestone roof. Chemical measurements of these two sources today show a $P\text{CO}_2$ of the diffuse infiltration to be unusually low – around 0.0006 atm. Small streams fed by recharge from the limestone

Figure 3. Intense corrosion by seepage of low-$P\text{CO}_2$ water into Mammoth Cave, Kentucky, after closed-system dissolution of limestone at the base of a sandstone cap-rock. The $P\text{CO}_2$ of the cave air is 0.0008 atm – very low for cave atmosphere. And yet the incoming seepage water readily absorbs CO_2 from the cave air and becomes highly aggressive.

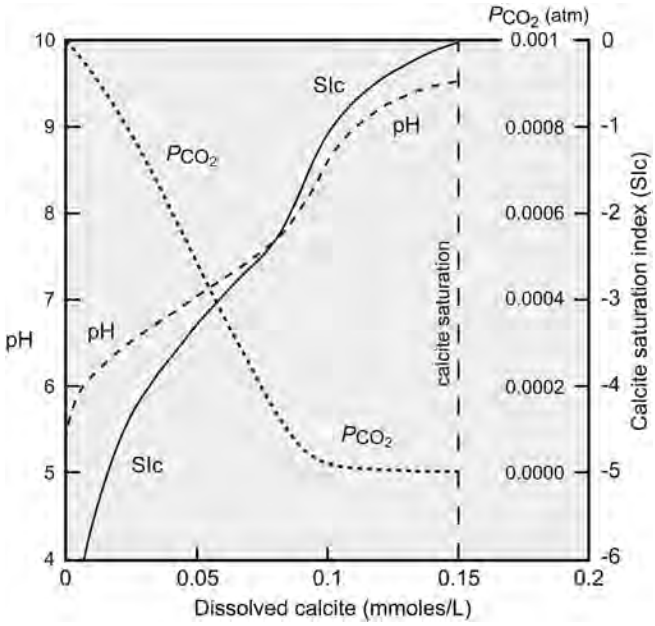

Figure 4. Evolution of water chemistry during closed-system dissolution of limestone. Note the extremely low PCO_2 and high pH of the water while it remains within the closed system.

outcrop have a higher PCO_2 of ~0.003 atm. PCO_2 of the cave atmosphere is ~0.00125 atm, according to analysis of standing pools water. (All analyses calculated from measurements of water chemistry in Back 2011 and Long et al., 2012.) Clearly, before the pH was measured, the diffuse infiltration water had gained CO_2 from the cave air, and the water from the outcrop had lost CO_2 to the cave air by degassing.

Mixing of these two waters would produce a solution undersaturated with calcite. Undersaturation would have been much greater if these waters had entered a water-filled cave where degassing into the atmosphere could not take place. If it were possible to measure the initial water sources (described above Figures 3 and 4), the amount of undersaturation would be about twice as great. The cave would be concentrated in the mixing zone and could not extend very far down-dip because the solutional aggressiveness would have been largely exhausted – as in mixing-zone caves in seacoast aquifers today.

Most of the present infiltrating water corrodes the cave ceiling, but it has little chance to mix with inflowing water from the limestone outcrop areas, as it did when the caves were presumably filled with water. Also, the water supply appears to be much less today than when the caves were actively enlarging, as shown by the large paleo-valleys up-slope from the caves.

Measurement of infiltration rate through the Minnelusa cap has been measured at both caves by Wiles (1992). They indicate an estimated 12,000 $m^3/yr/km^2$ of infiltration. Again, in today's semi-arid climate the recharge rate appears to be much less than in the past. The landscape has long been rather static, less so than today with its Pleistocene entrenchment. But even using

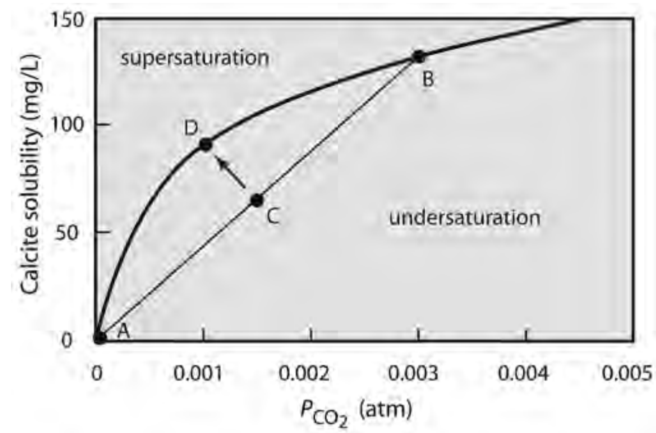

Figure 5. Effect of mixing corrosion between two waters similar to those entering Wind Cave today through the Minnelusa cap-rock (A), and from the Madison outcrop (B). Although both are at or near saturation at their respective PCO_2 values, the mixture (C) would be highly undersaturated if mixing took place rapidly. In reality, equilibrium is reached by a continuous migration toward D (following closed-system evolution where CO_2 is consumed as calcite dissolves). The overall dissolution rate is governed mainly by the rate of input of the smaller water source, probably the infiltrating water. Note the optimum effect caused by one source having virtually zero PCO_2.

Figure 6. Limestone blocks (embedded in paleokarst fill) are being corroded today by seepage water entering Wind Cave today through overlying sandstone and sandy sediment at the base of the Minnelusa Formation (Many similar-looking cupolas and domes are formed instead by condensation corrosion).

the present recharge rates, and water chemistry compromised by CO_2 exchange with the cave air, the caves could have been generated within 2 million years. If adjustments are made for the likely water chemistry when the caves were water-filled, the required time would be closer to 1 million years. Was this time span available? Keep in mind that the caves are adjusted to the present landscape and outcrop pattern, and that the thick calcite wall crusts were deposited in the late Oligocene, more than 20 million years ago. The landscape remained nearly static during speleogenesis and has not changed substantially since then.

CONCLUSIONS

The mixing model, unlikely as it might seem, avoids all of the pitfalls that trouble the other models. It accounts for the cave distribution, fits the geomorphic history, and, most importantly, it relies on geochemistry that is still valid today. There are still some unanswered questions about the nature of meteoric recharge when the caves were enlarging in the mid-Oligocene.

The speleogenetic water was not rising, so by some definitions this process would not be considered hypogenic (e.g., Klimchouk, 2007). Palmer (1991) defines hypogenic cave development as that where aggressiveness is generated by chemical processes below the land surface. By this definition, the mixing process envisioned for the Black Hills caves would qualify as hypogenic, and mixing-zone caves in seacoast aquifers would as well. Many of the features in caves of this origin are similar to those formed by rising water. There is no simple distinction between epigenic and hypogenic cave origin, and therefore any discussion of speleogenesis should include a description of the exact nature of the water source and its chemical aggressiveness.

There is more to the story. The initial sulfate beds within the limestone, together with the diagenetic and meteoric phases of early karst, were essential in preparing the way for Cenozoic enlargement of the caves to their present size. Their genesis also has some open-ended questions, but that topic is beyond the scope of this paper.

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ISOTOPIC STUDIES OF BYPRODUCTS OF HYPOGENE SPELEOGENESIS AND THEIR CONTRIBUTION TO THE GEOLOGIC EVOLUTION OF THE WESTERN UNITED STATES

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Hypogene speleogenesis in the western United States is associated with a deep source of water and gases that rise and mix with shallow aquifer water. Caves are formed below the surface without surface expressions (ie, sinkholes, sinking streams), and byproducts of speleogenesis are precipitated during the late phase of hypogene speleogenesis. These byproducts provide geochemical and geochronological evidence of a region's geologic history and include gypsum rinds and blocks, elemental sulfur, halloysite-10Å, alunite, natroalunite, and other sulfur-related minerals. The following speleogenetic and speleothemic features are common: alteration rinds, crusts, mammillaries, folia, rafts, and cave spar. The types of hypogene speleogenesis vary and many can be expressed in space and time in relation to paleo-water tables. We identify two general types: (1) H_2S-H_2SO_4-dominated speleogenesis that takes place predominantly near a paleo-water table (a few meters above and below), and (2) CO_2-dominated speleogenesis that mostly takes place 10s to 100s of meters below a paleo-water table, with latest-stage imprints within meters of the water table.

The Kane caves in Wyoming, and the Guadalupe Mountains caves in New Mexico and West Texas, are examples of H_2S-H_2SO_4-dominated speleogenesis (also known as sulfuric acid speleogenesis, SAS), where deposits of H_2S- and H_2SO_4-origin are the obvious fingerprints. The Grand Canyon caves in Arizona and Glenwood Caverns in Colorado are examples of CO_2-dominated systems, where H_2SO_4 likely played a smaller role (Onac et al., 2007). Deeper-seated geode-like caves, like the spar caves in the Delaware Basin area, are probably CO_2-dominated, and have formed at greater depths (~0.5 ± 0.3 km) below paleo-water tables. Caves in the Black Hills, South Dakota are composite and complex and show evidence for multiple phases of hypogene speleogenesis. In areas such as the Grand Canyon region, these paleo-water tables, when they existed in thick carbonate rock stratigraphy and especially at the top of the thick carbonate rock strata, were likely regionally relatively flat in the larger intact tectonic blocks.

Geochemical studies of these deposits are providing information about the timing of speleogenesis through U-Th, U-Pb, and Ar-dating. In addition, tracer data from isotopes of C, O, S, Sr, and U are indicators of the sources of water and gases involved in speleogenesis. From these studies, novel canyon incision and landscape evolution interpretations are appearing in the literature. Beyond this, the study of these byproduct materials seems to show evidence that the deeply sourced water and gases involved in hypogene speleogenesis in the western United States are generated during tectonic and volcanic activity, and may be related to mantle processes associated with formation of the Rocky Mountains, Colorado Plateau, Basin and Range province, and Rio Grande Rift.

INTRODUCTION

Hypogene speleogenesis involves caves that form 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" (Palmer and Palmer, 2000; Klimchouk, 2007). This is particularly true of many caves in the western United States (Palmer and Palmer, 2009). Hypogene speleogenesis in western United States caves results in various byproducts due to significant amounts of CO_2 and H_2S associated with the cave-forming processes (Hill, 1986; Polyak et al., 1998; Polyak and Provencio, 2001; Spilde et al., 2006; Palmer and Palmer, 2012). Most of these are listed in Table 1. These byproducts provide researchers the opportunity to better understand the hypogene speleogenetic processes through geochemical analyses. This seems to be particularly true for many of the western United States caves. Byproduct materials exist in caves from the Rocky Mountains, Colorado Plateau, Rio Grande Rift, and Basin and Range provinces. These materials archive within them multiple geologic "proxies." In other words, analyses of these materials are telling stories not only about speleogenesis, but about landscape evolution and mantle-processes. Researchers have only touched on the potential of the study of these materials.

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Figure 1. Selected hypogene cave areas in western North America. AC = travertine caves in Alberta, Canada, BH = Black Hills, GC = Grand Canyon, GcCo = Glenwood Canyon, GM = Guadalupe Mountains, , LC = Lehman Cave, nM = northern Mexico, nWy = Kane Caves, sAz = southcentral Arizona, and WCo = Cave of the Winds.

PREDOMINANTLY H_2S-H_2SO_4-DRIVEN SAS EXAMPLES

Earliest analyses on speleogenetic byproducts were the sulfur isotope measurements for gypsum blocks in Carlsbad Cavern in the Guadalupe Mountains adjacent to the Delaware Basin. Hill (1981), Kirkland (1982), and Hill (1987) showed with $\delta^{34}S$ measurements that the large quantities of gypsum in Carlsbad Cavern could not be marine-derived, but instead the gypsum was a byproduct of SAS, and that the speleogenesis involved H_2S that likely migrated from hydrocarbon deposits in the adjacent Delaware Basin. Van Everdingen et al. (1985) also described a similar process going on in small travertine caves in Banff National Park, Alberta, Canada, that they supported with $\delta^{34}S$ and $\delta^{18}O$ values in the sulfate minerals. In Guadalupe Mountains caves, another speleogenetic byproduct, alunite, was used to determine the timing of speleogenesis. The µm-sized pseudo-cube crystals of alunite were separated from a clay (halloysite-10Å) matrix by two independent methods (elutriation and HF-treatment), and then was analyzed by two different $^{40}Ar/^{39}Ar$ methods (capsulated and non-capsulated) (Polyak et al., 1998; Polyak et al., 2006). The results were consistent and showed the evolution and the timing of H_2S-H_2SO_4-related speleogenesis. In both isotope systems, it was the sulfur ingredient of speleogenesis that made this possible. The Kane caves, H_2S-H_2SO_4-caves identified by Egemeier (1981) in Wyoming are much smaller and younger than the H_2S-H_2SO_4-caves parallel to the Guadalupe reef escarpment. One of these caves has been used as a modern analog of this process generating substantial geomicrobiological interest (Engel et al., 2004). So far, the material and geochemical evidence seems to suggest that the H_2S-H_2SO_4-caves form a few meters below and above the water table. The cave data are also relevant to other fields; for example, theories for migration of hydrocarbons have come from $\delta^{34}S$ values of cave gypsum and alunite, and the timing of displacement of the Guadalupe tectonic block (or timing of water table descent relative to the tilted block) is made possible by the Ar-Ar ages of the alunite. The erosional history of the Guadalupe Mountains could be inferred because of the isotopic results of speleogenetic byproducts (DuChene and Martinez, 2000). More recently, $\delta^{13}C$ and $\delta^{18}O$ isotope values for "late stage" dolomite and calcite are being used to work out the terminal hypogene speleogenetic environment (Palmer and Palmer, 2012).

PREDOMINANTLY CO_2-DRIVEN EXAMPLES

The Grand Canyon hypogene caves have a H_2S-H_2SO_4 component to their speleogenesis, but the byproducts of terminal speleogenesis (spar linings, mammillaries, folia, and cave rafts) suggest that CO_2 was likely the most important ingredient (Hill and Polyak, 2010). The calcite spar linings (druses of euhedral calcite lining passage surfaces) probably had to form at least 10s of meters below the water table, but after the cave passages were formed at greater depths from the same rising gases and solutions. The mammillaries, folia, and rafts are evidence of the terminal stage of hypogene speleogenesis given that this process is linked to deep-seated CO_2. The mammillary calcite that formed within meters of and below the water table contains sufficient uranium concentrations to be dated using U-Pb analyses. Polyak et al. (2008) offered an interpretation of incision history for Grand Canyon based on the analyses of these terminal hypogene speleogenetic byproducts. The U-Th and U-Pb results of the mammillary calcite were interpreted to be a geologic proxy for incision rate history. U-Th results were used to make a contribution to the incision history of Glenwood Canyon several hundred km upstream of the Grand Canyon in Colorado. While there is evidence for SAS in caves of Grand and Glenwood canyons, spar linings, mammillaries, folia, and cave rafts seem to weigh heavier towards evidence of a CO_2–driven system as do studies of springs in these areas (Luiszer, 1996; Crossey et al., 2009).

Some caves, generally smaller in size and commonly geode-like, show evidence for deeper speleogenesis. The spar caves of the Guadalupe Mountains are lined with large euhedral crystals of calcite. These small caves and large crystals were assigned a low temperature hydrothermal origin with fluid inclusion results of the spar calcite, suggesting a range in temperature of 30 to 85 ºC (Lundberg et al., 2000). A U-Pb age of 92 ± 7 Ma was obtained suggesting that the Laramide was a time of heating and deeply circulating low temperature hydrothermal water, and that the age represents a time of uplift of the western United States including the Guadalupe Mountains region. These caves prob-

Table 1. List of speleogenetic byproducts in western USA hypogene caves.

Mineral	Formula	Cave Area
Cave minerals formed directly or indirectly from hypogene speleogenesis		
Aluminite	$Al_2(SO_4)(OH)_4 \bullet 7H_2O$	GM
Alunite	$KAl_3(SO_4)_2(OH)_6$	GM
Amorphous silica	$SiO_2 \bullet nH_2O$	GM
Barite	$BaSO_4$	GM, nM
Boltwoodite	$(K,Na)(UO_2)(SiO_3OH) \bullet 1.5H_2O$	nM
Calcite/aragonite	$CaCO_3$	all areas
Celestite	$SrSO_4$	GM
Chalcophanite	$ZnMn^{4+}{}_3O_7 \bullet 3H_2O$	sAZ
Copiapite	$Fe^{2+}Fe^{3+}4(SO_4)_6(OH)_2 \bullet 20H_2O$	GM
Coquimbite	$Fe^{3+}2(SO_4)_3 \bullet 9H_2O$	GM
Dolomite	$CaMg(CO_3)_2$	GM, BH
Fluorite	CaF_2	GM
Gibbsite	$Al(OH)_3$	GM
Goethite	$FeO(OH)$	GM, GC, nM
Gypsum	$CaSO_4 \bullet 2H_2O$	all areas
Halite	$NaCl$	GC
Hematite	Fe_2O_3	GC
Halloyhsite-10 Å	$Al_2Si_2O_5(OH)_4 \bullet 2H_2O$	GM
Hydrobasaluminite	$Al_4(SO_4)(OH)_{10} \bullet 15H_2O$	GM
Jarosite	$KFe^{3+}{}_3(SO_4)_2(OH)_6$	GM
Metatyuyamunite	$Ca(UO_2)_2(VO_4)_2 \bullet 3H_2O$	GM
Natroalunite	$NaAl_3(SO_4)_2(OH)_6$	GM
Nordstrandite	$Al(OH)_3$	GM
Quartz	SiO_2	GM, BH
Ranciéite	$(Ca,Mn^{2+})_{0.2}(Mn^{4+},Mn^{3+})O_2 \bullet 0.6H_2O$	GM
Römerite	$Fe_3(SO_4)_4 \bullet 14H_2O$	GM
Sulfur (elemental)	S	GM
Todorokite	$(Na,Ca, K, Ba, Sr)_{1-x}(Mn,Mg,Al)6O_{12} \bullet 3\text{-}4H_2O$	GM
Tyuyamunite	$Ca(UO_2)_2V_2O_8 \bullet 5\text{-}8H_2O$	GM
Bedrock Minerals		
Calcite CaCO3		
Crandallite-beudanite group	$AB_3(XO_4)_2(OH)_6$, where A=Ca, Ce, Sr, Th; B=Al, Fe; X=P, S, As	
Dickite/kaolinite	$Al_2Si_2O_5(OH)_4$	
Dolomite	$CaMg(CO_3)_2$	
Illite-group minerals	$K(Al,Mg,Fe)_2(Si,Al)_4O_{10}(OH)_2$	
Illite-smectite mixed-layers	$KA_{14}(Si,Al)_8O_{10}(OH)_4 \bullet 4H_2O$	
Montmorillonite	$(Na,Ca,K)_{0.3}(Al,Mg)_2Si_4O_{10}(OH)_2 \bullet nH_2O$	

Formulas use the IMA-CNMNC list of names. GM = Guadalupe Mountains caves, nM = northern Mexico cave, sAZ = central Arizona cave, BH = Black Hills caves, GC = Grand Canyon caves, GcCo= Glenwood Caverns, Colorado, Colorado, AC = Alberta, Canada caves, nWy = northern Wyoming caves.

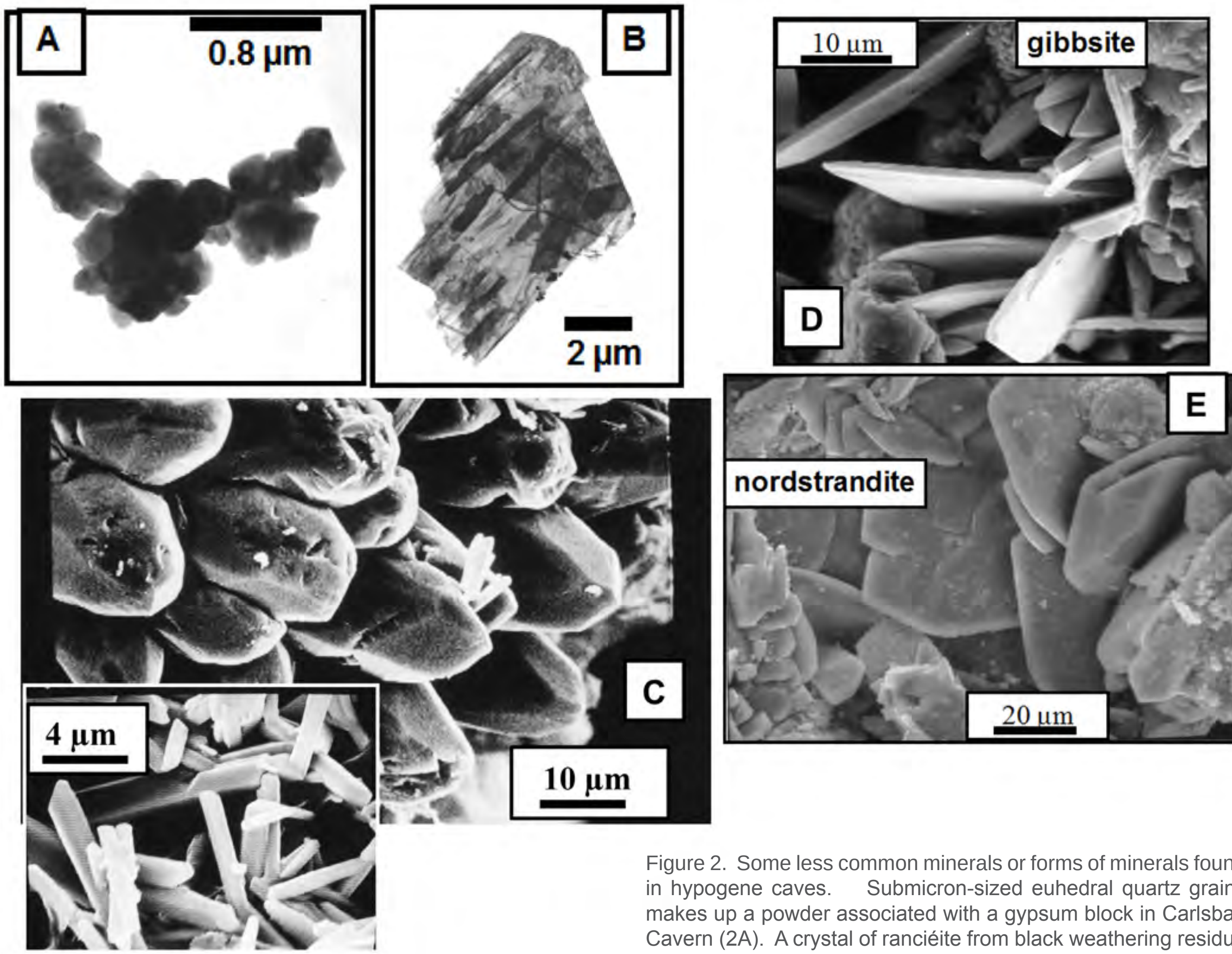

Figure 2. Some less common minerals or forms of minerals found in hypogene caves. Submicron-sized euhedral quartz grains makes up a powder associated with a gypsum block in Carlsbad Cavern (2A). A crystal of ranciéite from black weathering residue on calcite spar crystals (2B). Euhedral dolomite crystals and associated boltwoodite crystals in a gypsum crust also containing quartz (2C). Bladed crystals of gibbsite and nordstrandite associated with nodules of speleogenetic byproducts (2D and E).

ably formed from CO_2-rich waters rising from below, and they probably formed deep below the water table. U-Pb age of 232 Ma (Triassic) for spar from a small Grand Canyon spar cave provides evidence that this cave and those crystals formed roughly 750 m below the water table because the Triassic beds above the cave are near-marine or shallow marine sediments (Morales, 1990) where deep water tables probably could not have existed (Decker et al., 2011).

It is from this research that we ponder a definition for hypogene speleogenesis of these western USA caves that includes the speleothem assemblage of mammillaries, folia, rafts, and gypsum rinds from deep-seated sources of CO_2 and H_2S as evidence for speleogenesis. In other words, if it can be shown that the mammillaries of Devils Hole (a tectonic crack), for example, have the same sources of CO_2 as those of the hypogene caves in Grand Canyon, then even though a dissolution phase has not been demonstrated, is this evidence for a component of hypogene speleogenesis?

COMPLEX COMPOSITE EXAMPLES

Two of the world's longest caves are located in the Black Hills of South Dakota. These are Jewel Cave, in Jewel Cave National Monument, and Wind Cave, in Wind Cave National Park. Like the Guadalupe and Grand Canyon caves, they have a hypogene, albeit more complex origin. They contain abundant clues to the regional geologic history that date back to the Mississippian Period, soon after the host limestone was deposited (Palmer and Palmer, 2008). Determining the age of these deposits will contribute significantly to this long history of the caves.

The Late Mississippian karst was buried by 1–2 km of mainly clastic sediments of Pennsylvanian through Cretaceous age, and the early karst phases were preserved as true paleokarst. Voids not filled with sediment were lined by the dogtooth spar of presumed Mesozoic age, and like the Grand Canyon, these euhedral well-formed crystals represent deep-seated conditions. Also, in areas of both Jewel and Wind caves, walls are lined with coarse-

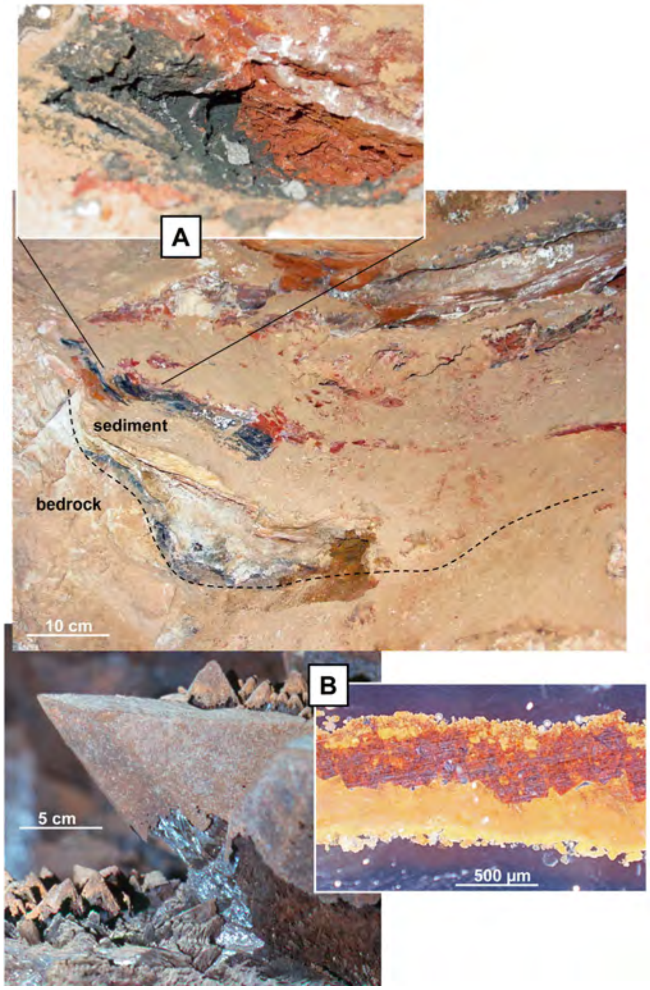

Figure 3. Fe- and Mn-rich sediment overlapping bedrock in a Grand Canyon cave (A). These materials are interpreted to be related to speleogenesis of this cave, although further study is needed to distinguish whether it is the product of cave genesis or the product of earlier breccia pipe mineralization. Cave hood on cave spar is ~1 mm thick (B). The crystalline material of this Grand Canyon cave hood consists of goethite. Reflected light image of a thin section shows two distinctly different colored reflections. These materials undoubtedly hold several potential geologic proxies.

ly crystalline calcite ("nailhead spar") averaging 15 cm thick. Similar to Grand Canyon caves, this is a palisade-like spar lining, often interpreted as the final stages of thermal cave development; and yet the spar overlies a weathered bedrock surface, indicating that the caves had drained prior to spar emplacement. Palmer and Palmer (1989) interpreted the nailhead spar to have formed during blockage of springs around the outer edges of the Black Hills during the Oligocene, with gradual draining of the caves as the continental sediments were removed by late Tertiary erosion. In 2008 we used U/Pb dating to find dates that range from 25 Ma at the base of the spar to 15 Ma at the outer edge (Palmer et al., 2009). This correlates perfectly with the highest levels of sediment accumulation during the Late Oligocene – Early Miocene, illustrating the importance of cave spar studies. This spar probably developed in relatively shallow conditions, at depths of about 150 to 300 m. Similar to the Grand Canyon, the palisade-like spar linings formed in shallower conditions (10s to a few hundreds of meters), while the older, well-formed scalenohedral/rhombohedral crystals formed in deeper environments.

REGIONAL WATER-TABLE PROXY

The byproducts of hypogene speleogenesis offer speleologists powerful new geologic proxies. Absolute timing of speleogenetic events is critical to these studies. U-Th, U-Pb, and $^{40}Ar/^{39}Ar$ are the dating methods most applied to dateable minerals, and potential minerals for dating include calcite, aragonite, dolomite, quartz/opal, Mn- and Fe-oxides, and gypsum. The hypogene features can also be related to a paleo-water table or potentiometric surface. The cave-forming component of CO_2-driven hypogene speleogenesis occurs where CO_2 is prevented from degassing by depth below a water table (unconfined) or potentiometric surface (confined), but the latest phase of this speleogenesis (deposition of byproducts) happens near or at the water table. Polyak et al. (2008) used results from analyses of these deposits in Grand Canyon to define paleo-water table positions, assuming relatively flat water tables in the thick Muav-Redwall carbonate sequence (massif). Their assumption is the introduction of a general concept that "the least likely place to elevate or perch a water table is at the top of a karst aquifer" seems reasonable in areas where the thick carbonate rocks are not too broken up as is/was the case for the Grand Canyon area. So the concept introduced by Polyak et al. (2008) is that the folia, mammillaries, rafts, and gypsum rinds represent the paleo-water tables, and the paleo-water tables represented the ancestral Colorado River levels. In the case of H_2S-H_2SO_4-hypogene caves, speleogenesis seems more likely to take place near a water table, and therefore the cave passages represent paleo-water tables. The deep spar caves and the spar that lines them form 100s of meters below the water table. In each case, paleo-water tables or paleo-potentiometric surfaces, through the understanding of hypogene speleogenesis, become a tool for reconstruction of landscapes or canyon histories.

DISCUSSION AND CONCLUSIONS

Additional evidence coming from geochemical and isotopic studies is particularly useful. $\delta^{13}C$, $\delta^{18}O$, $\delta^{34}S$, $\delta^{88}Sr$, $^{87}Sr/^{86}Sr$, $\delta^{234}U$, and $\delta^{238}U$ are already applicable as isotope tracers for the hypogene speleogenesis systems. Each or combinations of them are useable as geologic proxies. For instance, $\delta^{18}O$ and $\delta^{34}S$ of gypsum are useful together to more robustly distinguish H_2S-H_2SO_4-hypogene speleogenesis from other speleogenetic processes (Onac et al., 2011). $^{87}Sr/^{86}Sr$ isotope values are effective tracers of fluid sources in hypogene speleogenesis systems. $\delta^{88}Sr$ values of calcite may yield temperature information, especially where fluid inclusion analyses are not practical (Fietzke and Eisenhauer, 2006). $\delta^{238}U$ values may be used to distinguish hypogene byproducts from vadose speleothems in some regions. Some minerals in the assemblage listed in Table 1 and associated with hypogene caves, have yet to be studied in detail. Figure 2 shows a few of the less common minerals. Each will have isotopic systems that will undoubtedly contribute more proxies. Mn- and Fe-oxides are more common (Figs. 3 and 4), and hold the same potential. Some Mn- and Fe-oxides are dateable by (U-Th)/He and Ar-Ar methods (Shuster et al., 2005).

Thus far, these tools are telling us that the sources of solutions and gases coming from depth and mixing with aquifer groundwaters to create hypogene speleogenesis in the western United States are related to volcanism/tectonism since the Laramide orogeny (with a few examples of older events). These caves and deposits have been formed in the Basin and Range provinces from Nevada to Texas and northern Mexico; in the Rio Grande Rift province in southern New Mexico and northern Mexico (Hill, 1987; Lueth et al., 2005); along the margins of the Colorado Plateau in Grand Canyon, Arizona, and Glenwood Canyon, Colorado; in the Black Hills of South Dakota; and in the Rocky Mountain provinces (Luiszer, 1996). The timing of mammillary formation in Grand Canyon seems to coincide with important volcanic or tectonic events. The idea is that volcanic events can serve as "pumps" that force these hypogene solutions and gases upward (increase deep circulation); and that episodes of faulting open up joints that allow these deep solutions to move upward. Mantle processes are probably the drivers of the volcanic and tectonic activity.

Figure 4. Stalactitic-like deposits of Mn- and Fe-rich materials coated with mammillary calcite. Note the black material in the broken piece on the cave floor in the upper photo. The deposit of Mn- and Fe-rich material under the mammillary coating shows that this material is older than the mammillary calcite. Even though there was too much common Pb in the mammillary to allow U-Pb dating, its position in the canyon suggests that it is roughly 2 Ma, indicating that the underlying black deposit is older.

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DEVILS HOLE: ANTIKARST IN THE MOJAVE

Alan C. Riggs[1]

The "classic" view of thermal hypogene caves is that rising (artesian?) warm waters become aggressive as they cool, dissolving carbonate to produce 3D caves with particular morphologies, such as mazes, spongework, and upward-pointing dissolution morphs. Devils Hole, Nevada, is a cave where rising 33-34°C waters precipitate thick coatings of mammillary (phreatic) calcite on the walls of a simple planar (2D), nearly vertical fissure. Why the difference?

To understand Devils Hole's contrary behavior, we need a basic understanding of the Ash Meadows groundwater flow system that feeds Devils Hole and the adjacent Ash Meadows spring line. The Ash Meadows flow system (which underlies a land area of at least 12,000 km^2) is a sub-basin of the Death Valley regional flow system, which drains a huge area of the southern Great Basin to the ultimate sink in North America, Death Valley. Climatically, the area of the Ash Meadows flow system is Mojave Desert, with average annual precipitation of 10 cm or less on the basin floors (<1000 m elevation), and possibly as much as 1 m at the highest elevations (roughly 3,600 m elevation) of the Spring Range west of Las Vegas, Nevada. The geology of the Ash Meadows flow system is varied, but is dominated by thick (averaging ~4,600 m) deposits of Paleozoic marine carbonate rocks, which, in places, extend from the basal aquitard to the ridge tops, are widely exposed at the surface, and underlie more recent alluvial and volcanic deposits in other areas. The carbonate rocks are contiguous throughout much of the area of the flow system; given their brittleness and long history of tectonic deformation (up to and including modern Basin and Range extension), they host extensive networks of interconnected open fractures. The fracture networks act as integrated highly transmissive conduit systems that transport predominantly high-elevation recharge to basin-floor discharge points. At the Ash Meadows flow system's downstream end, the fracture-flow network is faulted up against a poorly transmissive aquitard, which causes the groundwater to well to the surface and discharge as the 16-km-long Ash Meadows spring line along the eastern flank of the Amargosa Desert.

For the purposes of this discussion, the geneology of the Ash Meadows flow system can be divided into three parts. The recharge areas are classic epigene karst where infiltrating precipitation becomes aggressive as it dissolves CO_2 in the thin soil zone, and then dissolves carbonate bedrock deeper underground. Given the present desert climate, epigene karstification is poorly developed. As the water infiltrates deeper, it eventually recharges the aquifer and circulates through the fracture network to at least 1 km below land surface; in this part of the flow system, flow is more lateral than vertical and the regime could reasonably be called tectonogene, i.e. the primary volume generator is the continuing opening of conduits by tectonic extension. Hence, speleogenesis at depth is hypogene in the dictionary sense – that is, formed/generated/born at depth below land surface. This use of hypogene is at odds with the "classic" use of that term, so tectonogene may be a better descriptor, and does not conflict with any other conventional usage. The lateral-flow tectonogene part of the flow system continues until groundwater flow approaches the aquitard barrier, at which point the water rises to discharge along the Ash Meadows spring line. The water level in Devils Hole, though about 50 m above the spring pool elevations in Ash Meadows is about 17 m below the land surface, making Devils Hole, in effect, a tectonic cenote. The configuration of the rising part of the aquifer beneath Devils Hole hydrologically resembles "classic" thermal hypogene regimes in that it is inundated by warm water rising from depth.

However, the morphology of Devils Hole to ~150m below the water table is very different than classic thermal hypogene morphologies. The Devils Hole fissure is an open nearly vertical 2D fault plane; the water in it, instead of becoming aggressive as it rises and cools, becomes slightly supersaturated with respect to calcite, turning the rising limb into an antikarstic regime (i.e. a regime where rock precipitation is a primary geomorphic process) that precipitates mammillary calcite on all phreatic surfaces -- and has done so for at least a half a million years! How can that be???

The answer seems to be that even though the PCO_2 of the deep Ash Meadows flow system water is sufficiently high (roughly 10-2 atm) for the water to become aggressive (undersaturated) on cooling, it apparently rises sufficiently rapidly that saturation decrease due to cooling is more than compensated by saturation increases resulting from pressure release: water saturated wrt calcite at 1 km depth becomes oversaturated by about 58 µmoles per kilogram when it is lifted to surface pressure at constant temperature. That parcel of water would have to cool by more than 7 °C, to compensate for the pressure-release-derived supersaturation.

Given that the PCO_2 of Devils Hole water is about 40 times atmospheric, it might be argued that supersaturation in Devils Hole water is abetted by, or even primarily caused by, outgassing. The presence of modern-day rafts and folia in Devils Hole unambiguously shows that CO_2 outgassing from the water surface can stimulate rapid carbonate precipitation, but there are several reasons for be-

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lieving that outgassing is not necessarily the primary, or even a significant contributor to mammillary (phreatic) calcite precipitation. First, unlike some cave pools where CO_2 outgassing has been interpreted as fostering precipitation of a carbonate girdle around the pool shallows, there is no precipitation on the deeper bare bedrock surfaces below the girdle. The mammillary coating on the underwater surfaces in Devils Hole extends to at least 150 m below the water table, and its thickness does not noticeably diminish with depth or lateral distance from the entrance. Second, there has not always been a Devils Hole. Prior to a flow conduit stoping to the surface to form Devils Hole, it was roofed over with bedrock. The pre-collapse semi-closed system would have restricted CO_2 outgassing relative to today's configuration, presumably with similar reductions in the component of mammillary precipitation attributable to outgassing. Third, mamillary precipitation tends to coat all phreatic surfaces with a uniformly thick layer of calcite, progressively filling and sealing all cracks and holes, with the effect of further reducing the escape of CO_2 and hence outgassing-driven speleothem precipitation. In fact, given enough time, mammillary precipitation would eventually fill and plug the conduits comprising the discharge end of the aquifer, driving the discharge area up-gradient. A 10-m-wide mammillary vein exposed at land surface in eastern Nevada, shows that, given enough time, mammillary calcite will fill even broad fissures. On the other hand, in the Ash Meadows flow system, the tendency for conduits to be sealed and filled with mammillary calcite is offset by ongoing fissure widening, which adds volume faster than mammillary calcite can fill it. A consequence of this is that the mammillary seals are broken with each bout of extension, presumably allowing a new burst of CO_2 outgassing that is progressively throttled back by continuing mammillary overgrowth and sealing, until the next time the conduit widens a bit more, and the cycle repeats. That mammillary precipitation occurs in the absence of outgassing is suggested by the presence of numerous calcite veins in cemented fanglomerate cliffs above Furnace Creek Wash, a tributary to Death Valley from the east. At least one of the calcite veins (as mammillary-lined or -filled fractures too narrow to enter are traditionally called) is easily traceable tens of meters up the cliff face. Near the top, the vein expands into a big carbonate mound – a fossil spring mound. In desert environments, spring mounds are hot beds of biological activity that, like soils, tend to be zones of high PCO_2. Hence, calcite precipitation in the vein that fed the fossil spring mound is unlikely to have been augmented by outgassing.

So there you have it; the Devils Hole antikarst rises to the challenge of showing that there is more than one way to be hypogene.

Acknowledgments: David Parkhurst used Phreeqc to calculate the calcite solubilities at different temperatures and pressures used herein.

HYPOGENE PALEOKARST IN THE TRIASSIC OF THE DOLOMITES (NORTHERN ITALY)

Alberto Riva[1]

In the Triassic of successions of the Italian Dolomites (Northern Italy), there are several examples of different types of hypogene paleokarst, sometimes associated with sulfur or hematite ore deposits.

The paleokarst features are related to a regional volcanic event occurred during the Ladinian (Middle Triassic) that affected several carbonate platforms of Anisian-Ladinian age.

This study is focusing mainly on the Latemar paleokarst, in the Western Dolomites, and on the Salafossa area in the Easternmost Dolomites.

The karst at Latemar developed as the result of a magmatic intrusion located just below the isolated carbonate platform, developing a system of phreatic conduits and some underground chambers, not justified by the entity of the submarine exposure occurring at the top of the carbonate platform. Most of these features are located about 500 m below the subaerial unconformity and are filled with middle Triassic lavas. Only in one case, the filling is represented by banded crusts now totally dolomitized, with abundant hematite. In this case, the only way to explain the presence of the karst at this depth is to invoke a deep CO_2 source allowing the dissolution of the carbonate at such depths: the fact that some phreatic conduits and a possible underground chamber are filled only with lavas is pointing toward an important role of volcanism in karst development.

Salafossa is a well-known mine located in the easternmost Dolomites and has been exploited until 1986, when all the activity ceased. The main metals, in this case, are Zn-Pb-Ba-Fe, exploited within a quite complex paleokarst system developed in several levels, filled by a complex mineralized sequence. The strong dissolution led to the development of voids aligned with the main fault controlling the mineralization, with a proper karst system with phreatic morphologies.

INTRODUCTION

The Italian Dolomites are a type area for the Triassic period, well studied since the 19th century, but are still preserving some good examples of geological phenomena, including hypogene paleokarsts of Ladinian age.

The presence of mineralization associated with karst in the Dolomites is known since the late 1960s, when a detailed study was performed on the Salafossa Mine in the Easternmost Dolomites, but, in general, the Triassic paleokarsts have been never directly associated with hydrothermal fluids, but only to subaerial exposures (Cros and Lagny, 1969)

In this paper, two striking examples are shown, one in the western (Latemar) and one in the easternmost Dolomites (Salafossa) (Figure 1). There are also some small other examples in the area, but usually are of minimal extent.

Both these examples are related to a regional volcanic event, occurred during Ladinian, that led to an important magmatic intrusion close to Latemar (Predazzo) and also triggering hydrothermal circulation creating several mineral deposits, sometimes associated to paleokarst.

LATEMAR MASSIF

The Latemar platform is a well known Middle Triassic (late Anisian) isolated carbonate buildup, cropping out in the western Dolomites, northern Italy. This carbonate buildup has a diameter of about 4 km and is characterized by a flat bedded platform interior (Latemar Limestone) surrounded by steep (35°) megabreccia slope deposits (Sciliar Formation). This isolated platform developed in a very short time span, due to exceptional subsidence rates. Below the isolated buildup, another Anisian carbonate platform (Contrin Fm) is in continuity, separated by a sequence boundary completely masked by the dolomitization. The overall thickness of the two Anisian carbonate platform is close to 800 m (see Preto et al., 2011; Gramigna et al., 2013, and references therein).

The development of this carbonate platform was abruptly interrupted by a strong volcanic event, with a development of a magmatic intrusion below part of the platform and more eastward, called the Predazzo pluton. The calderic collapse related to this pluton crosscut a portion of the Latemar carbonate platform, with a collapse of more than 1500 m. The remaining portion of the carbonate platform was heavily intruded by more than 1600 volcanic dikes (Riva and Stefani, 2003).

Before the intrusion of the volcanic dikes, the platform was also affected by an hydrothermal dolomitization event, with complex pattern, mainly related to the geometry of the aquifer (Newton et al., 1990; Carmichael and Ferry, 2008; Carmichael et al., 2008).

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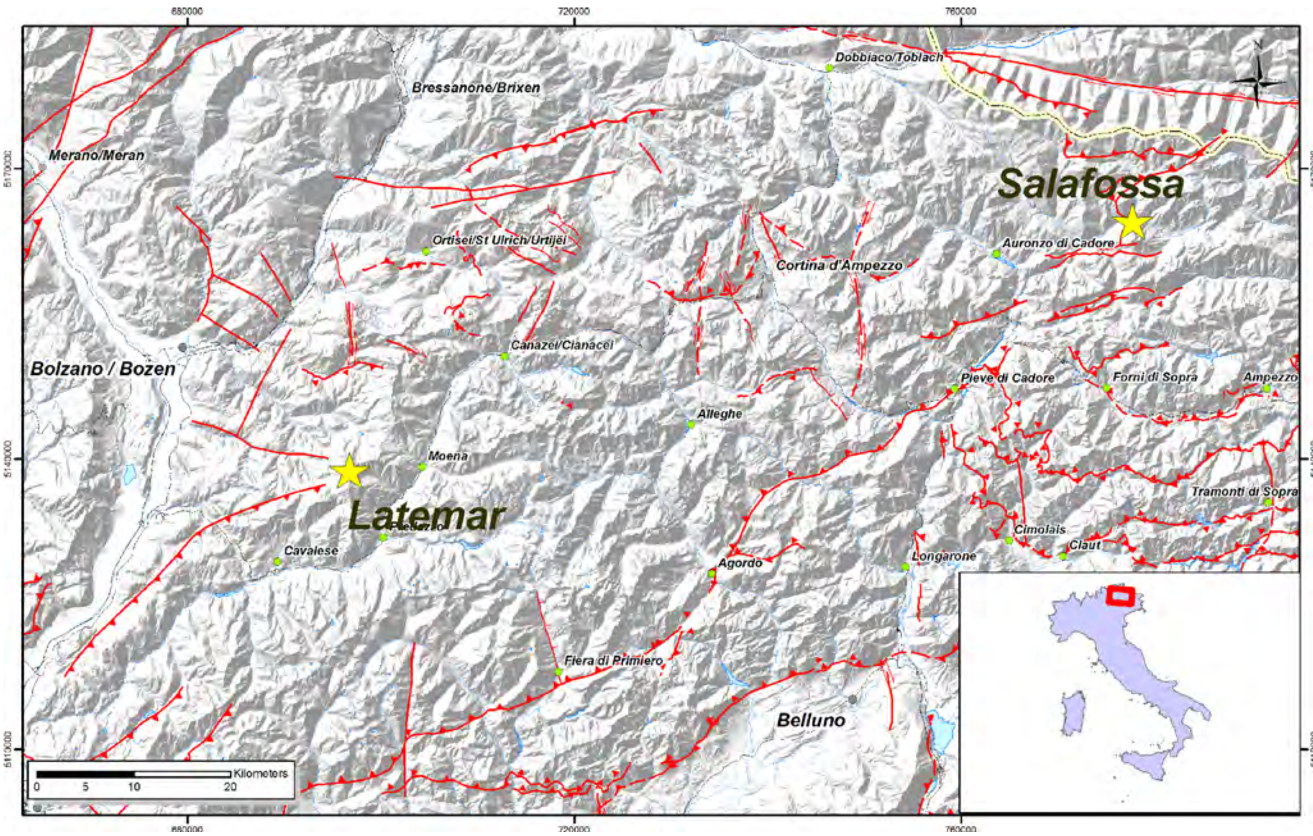

Figure 1. Location of the studied areas.

The Latemar Paleokarst Features

The Latemar carbonate platform very probably underwent a relatively short phase of subaerial exposure, related to the bulging effect of the emplacement of the Predazzo intrusion. Based on the available biostratigraphic data, the interval occurring between the bulging and the start of the volcanism was probably extremely short, below the biostratigraphical resolution of Triassic ammonoids (<500.000 years).

Some paleokarst features of the Latemar have been described by Cros and Lagny (1969), but they were always interpreted as associated to subaerial exposures and, in some case, they misinterpreted some volcanic diatremes as karst morphologies.

In the lower part of the platform, about 600 m below the platform top, there is a red and yellowish dolomitized feature particularly rich in hematite. The mineralized deposit is encased in dolomitic rocks, with a very sharp contact and covers an area of about 10m x 15m on a steep cliff. The filling appears to be quite complex, but from the base to the center there are 1) laminated dolomite/haematite alternations subparallel to the encasing rock surface resembling a flowstone with some intercalations of coarser material, 2) laminated red dolomite, 3) red and yellow dolomitized breccia with flat clasts and abundant vugs between the clasts, only partially filled with dolomite.

The encased mineralizations are totally different from the dolomites that are encased. The interpretation for this feature is a deep karst chamber, probably filled with flowstone crusts, then heavily dolomitized and mineralized. The position of this feature in respect of the platform top rules out a possible epigenic (hypergenic) origin, pointing more to a hypogenic origin.

Another feature is located about 500 m below the platform top and is represented by an horizontal subcircular conduit with a diameter of about one meter, cut by erosion, and completely filled with andesitic lava. The position of this feature, coupled by the morphology, could be explained only by deep karst related in some way to the hydrothermal fluids preceding the main volcanic phase.

In other areas, there is a peculiar feature, called “Le Porte Nere” (The Black Doors): it is a black subcircular feature on a slope with a dip of about 65°. This feature, very difficult to access, is showing a very irregular surface of contact with the encasing rock and the filling is composed exclusively by lavas. The shape of the encasing rock is very irregular, with a low dip base, and it is not cylindrical, ruling out the explanation of this feature as a diatreme. This feature could be explained as a karst feature, completely filled by lava: the distance from the platform top, however, is only 100 m, possibly pointing also to a epigenic origin of the void. However, the very short time span between the possible exposure event triggering the epigenic karst devel-

Figure 2. Phreatic conduit filled with lava on the Latemar Mountain.

opment and the volcanism could be easily explained with CO_2 enhanced karstification as in the previous features.

In the same area, there are some narrow subvertical breccia bodies with limestone clasts floating in a red matrix, sometimes associated with ferruginous oolites, located very close to the platform top.

The paleokarst development in this area is considered to be intimately associated with the volcanism and similar dissolutional features have been identified also in the Agnello carbonate platform, located few kilometers south of the Latemar. Specifically, at the very base of the carbonate platform, there are some subvertical fractures surrounded by diffuse porosity with vugs deriving by the dissolutional enlargement of microfractures, with pores up to 3 cm of diameter.

THE SALAFOSSA PALEOKARST SYSTEM

The Salafossa mine is a well-studied example of mineralized paleokarst of Triassic age, always ascribed to an exposure event (Cros and Lagny, 1969; Lagny, 1969; 1974; 1975), exploited as a Zn-Pb-Ba-Fe deposit until 1986.

The ore body, that correspond to the paleokarst infilling, has an approximate size of 700×200×100 m and is hosted within the Contrin Formation of Anisian age, also outcropping on the Latemar massif, overlain by the Sciliar Formation of Anisian-Ladinian age.

The lower part of the ore deposits is considered to be within a proper karst system, with horizontal conduits filled with different generations of minerals. The size of the dissolved conduits is often higher than 50 cm, now completely filled.

The mineralization event is considered to be coeval to the Ladinian volcanic phase in the Dolomites, with the development of a regional hydrothermal system linked to an increase of the heat flow. At that time, the mineralized zone was quite deep in respect to the possible exposure surface (>500 m), pointing to the development of an hypogene karst system subsequently mineralized. In a paper by Lagny and Michel (2003), they cite unpublished research by J.P. Devigne from 1979 that identified microcrystals associated filaments of bacterial origin in the lead sulphides of the Salafossa mine.

CONCLUSIONS

In the Italian Dolomites, many paleokarst features of Triassic age were considered as the result of exposure events related to sea-level lowstands. However, both Latemar and Salafossa paleokarsts developed at a depth of more than 500 m below surface (probably very close to sea-level): in these cases the paleo-hydrostratigraphic positions are incompatible with proper epigenic (hypergenic) karst and are more peculiar of hypogenic karst.

These are only two possible examples in the area, but there is the possibility that many other mineral deposits in the area are connected to hypogene paleokarst systems.

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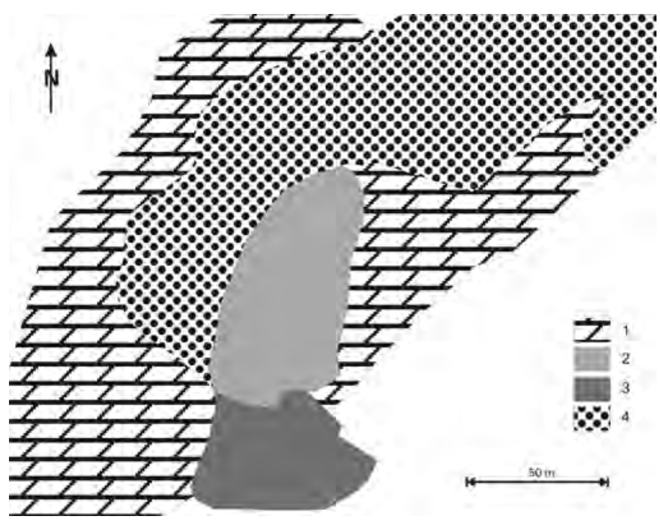

Figure 3. Sketch of the level 1154 of the Salafossa Mine (modified from Lagny 1974) with one of the largest paleokarst cavities (more than 50 m wide). The Anisian dolomite (1) is hosting a large mineralized paleokarst cavity (2) that passes to a brecciated zone (4). To the south, the cavity stops against some pyrite-rich dolomites (3).

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SPELEOGENESIS BY THE SULFIDIC SPRINGS AT NORTHERN SIERRA DE CHIAPAS, MEXICO, BASED ON THEIR WATER CHEMISTRY

Laura Rosales-Lagarde[1,2] and Penelope J. Boston[1,2]

Conspicuous brackish sulfidic springs have been described at the northern Sierra the Chiapas, Mexico. These springs are produced by a mixture between regional and local groundwater flow paths. The regional groundwater has an average Total Dissolved Ions of 3081 mg/L so it has a brackish composition. This brackish water is saturated with respect to calcite and dolomite but undersaturated with respect to gypsum, anhydrite and halite. The mass balance and the discharge rate are used to quantify the mass and volume of minerals that are dissolved by the brackish spring water following Appelo and Postma (1993). This quantification will allow comparing the various speleogenetic mechanisms in the area. This is considering the composition of the spring water is relatively constant over time, as it is suggested by periodic measurements at the Cueva de Villa Luz springs during the last 10 years.

Sulfur isotopes in the water are consistent with anhydrite dissolution as the main source of the sulfate to the brackish spring water. Thus, the average 6 mol/L of sulfate in the brackish springs are produced by dissolution of 6 mol of anhydrite after subtracting the sulfate that could result from evapotranspiration of rainwater. Each liter of brackish water dissolved an average of 882 mg of anhydrite, which are equivalent to dissolving 0.36 cm^3 of this mineral considering a density of 2.981 g/cm^3. Additionally, using the average brackish water discharge rate of 144 L/s, an average of 57 g of anhydrite are being dissolved each second per every liter of brackish water. This is a minimal value because some of the sulfate in the water is used by sulfate-reducing bacteria in the subsurface to produce the hydrogen sulfide in the spring water. The anhydrite subject to dissolution is found interbedded in the Cretaceous carbonates, either from the subsurface at 4,000 m below sea level to the carbonate outcrops.

Similarly, we can calculate the volume of halite that is being dissolved by the brackish springs, considering chloride is a conservative element and subtracting the chloride concentration from the rainwater from that of the spring water following Appelo & Postma (1993). The 22 mol/L of chloride in the brackish water can result from dissolution in the subsurface of 22 moles or 1.3 g of halite per liter of brackish water. This mass of halite dissolved is equal to 0.59 cm^3 considering a density of 2.168 g/cm^3. Alternatively, 118 g of halite are dissolved per second per each liter of brackish water if we use the average discharge rate of 144 L/s.

Even when the brackish springs are oversaturated with respect to calcite and dolomite, their dissolution is still possible due to the common ion-effect of calcium after anhydrite dissolution and by mixing of waters with different compositions. A range of 10 to 80 % of brackish water from the regional aquifers mixes with fresh water from the local aquifer based on their water chemistry. Additionally, sulfuric acid speleogenesis occurs due to the oxidation of hydrogen sulfide to sulfuric acid.

Finally, the increase in the chloride concentration of the fresh water springs with respect to the concentration in rainwater was used to estimate that from the 4000 mm/y of annual precipitation, only 4%, 158 to 182 mm/y, recharge the aquifers. This low percentage is slightly higher than the 3.3% recharge in marls, marly limestone, silts and clays (Sanz et al., 2011), probably because of the relatively small area of carbonate outcrops over the entire region and the lack of recharge in altitudes higher than 1500 m above sea level.

Sulfuric acid is the most obvious speleogenetic mechanism occurring in the caves of the northern Sierra de Chiapas, Mexico due to the high hydrogen sulfide concentration in the spring water. In addition, the location of the springs at a zone of regional and local discharge where waters from different composition converge and mix, and the amount of mixing calculated suggests mixing is also an important speleogenetic mechanism. However, the depth and the time constrains at which these two hypogenic mechanisms occur is still unknown. The relatively low rainwater recharge rate suggests epigenesis is limited. Most likely, the porosity created by dissolution of anhydrite and halite in the subsurface is occluded by the precipitation of calcite. Chemical modeling and petrography will help to elucidate the order of the reactions occurring in the subsurface.

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PONDERING THE IMPORTANCE OF SUBAERIAL CORROSION AS A SPELEOGENETIC AGENT

Ira D. Sasowsky[1]

Subaerial corrosion has been recognized as an important cave modifying process in limited settings. But is it possible that we overlook its importance in other cases? Could it actually be a significant speleogenetic agent in its own right? Numerous corroding agents have been identified including sulfuric acid, carbonic acid, ambient water vapor, and thermal water vapor. Morphogenetic features have been described, and cautions issued about possible confusion with hypogene features. Theoretical calculations seem to limit the importance of corrosion in many settings, but it appears that great care must be taken, especially for possible confusion between "hypogene" morphologies in a cave.

Some caves in the Iberian Range (Spain) seem undoubtedly hypogene in origin based on hydrologic constraints. They also contain morphologies that are consistent with this origin. But, extreme corrosion of speleothems and bedrock may be masking the nature of the cave morphology post-drainage of the forming waters. Topographic position of some caves suggests the possibility of a strong component of subaerial corrosion as the cave forming agent.

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FINGERPRINTING WATER-ROCK INTERACTION IN HYPOGENE SPELEOGENESIS: POTENTIAL AND LIMITATIONS OF ISOTOPIC DEPTH-PROFILING

Christoph Spötl[1] and Yuri Dublyansky[1]

Dissolution processes in karst regions commonly involve (meteoric) water whose stable isotopic (O, H, C) composition is distinctly different from that of the paleowaters from which the host rock (limestone, dolostone) formed. This, in theory, should lead to isotopic alteration of the host rock beyond the active solution surface as the modern karst water is out of isotopic equilibrium with the carbonate rock. No such alteration has been reported, however, in epigenetic karst systems. In contrast, isotopic alteration, commonly referred to as isotopic halos or fronts, are known from various hypogene systems (ore deposits, active hydrothermal systems, etc.). These empirical observations suggest that stable isotope data may be a diagnostic tool to identify hypogene water-rock interactions particularly in cave systems whose origin is ambiguous.

We have been testing the applicability of this assumption to karst settings by studying the isotopic composition of carbonate host rocks in a variety of caves showing clear-cut hypogene morphologies. Cores drilled into the walls of cave chambers and galleries were studied petrographically and the C and O isotope composition was analyzed along these cores, which typically reached a depth of 0.5 to 1.2 m. We identified three scenarios: (a) no isotopic alteration, (b) a sigmoidal isotope front within a few centimeters of the cave wall, and (c) pervasive isotope alteration throughout the entire core length. Type (a) was found in caves where the rate of cave wall retreat apparently outpaced the rate of isotopic alteration of the wall rock (which is typical, for example, for sulfuric acid speleogenesis). Type (c) was observed in geologically young, porous limestone showing evidence of alteration zones up to 5 m wide. The intermediate type (b) was identified in hypogene karst cavities developed in tight limestone, dolostone and marble.

Our data in conjunction with evidence from speleothems and their geochemical and fluid-inclusion composition suggest that the spatial extent of the isotopic alteration front depends on the porosity and permeability, as well as on the saturation state of the water. Wider alteration zones primarily reflect a higher permeability. Shifts are most distinct for oxygen isotopes and less so for carbon, whereby the amplitude depends on a number of variables, including the isotopic composition of unaltered host rock, the isotopic composition of the paleofluid, the temperature, the water/rock ratio, the surface of water-rock contact, the permeability of the rock, and the time available for isotope exchange. If the other parameters can be reasonably constrained, then semi-quantitative temperature estimates of the paleowater can be obtained assuming isotopic equilibrium conditions.

If preserved (scenarios b and c), alteration fronts are a strong evidence of hypogene speleogenesis, and, in conjunction with hypogene precipitates, allow to fingerprint the isotopic and physical parameters of the altering paleofluid. The reverse conclusion is not valid, however; i.e. the lack of evidence of isotopic alteration of the cave wall rock cannot be used to rule out hypogene paleo-water-rock interaction.

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